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HYDROTHERMAL ALTERATION AT SANTA RITA, NEW MEXICO. By Paul F. Kerr, J. L. Kulp, C. Meade Patterson, and Robert J. Wright

LATE CENOZOIC BLOCK FAULTING IN WESTERN MONTANA. By J. T. Pardee

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This paper is accompanied by a colored map. Separate copies of the map (unfolded, in tube) may be purchased. Remittance (25 cents per copy) should accompany orders.

STRATIGRAPHY AND THE STUDY OF METAMORPHIC ROCKS. By Marland P. Billings

TECTONIC THEORY VIEWED FROM THE BASIN RANGES. By Chester R. Longwell

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CENOZOIC GRAVELS ON THE RIM OF SYCAMORE CANYON, ARIZONA. By William E. Price, Jr.

NEW PROJECT GRANTS

J. Harlan Johnson, F.G.S.A., will study types of fossil coralline algae in museums abroad

Kazuo Kuroda will study the distribution of fluorine and chlorine in igneous rocks

J. Laurence Kulp, M.G.S.A., will establish a Strontium Isotope Method to measure age of sedimentary and metamorphic rocks

. Webb

C. Meade

AMPSHIRE

in tubes

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dimentary

LEGEND

STORM KING GRANITE

DISTRIBUTION OF GRANITE OUTSIDE BEAR MOUNTAIN REGION BASED ON AUTHOR'S RECONNAISSANCE EXCEPT IN WEST POINT QUADRANGLE NORTH OF LAT. 41°18' (AFTER BERKEY AND RICE, 1919).

BEAR MOUNTAIN REGION
COVERED BY THIS REPORT

CONTACTS, known and inferred

THRUST FAULTS, known and inferred, overthrust toward northwest
TRENDS OF KNOWN FAULTS IN WEST POINT QUADRANGLE AFTER BERKEY & RICE, 1919.

NORMAL (OR TEAR) FAULTS, known and inferred, with relative displacement

X———X' LINE OF SECTION OF FIG. 11

CONTOUR INTERVAL 20 FT., EXCEPT 100 FT. SOUTH OF LATITUDE 41°17'45"

1 1/2

BASE MAP AND GEOGRAPHIC NAMES FROM 'WEST POINT AND VIC

HUDSON HIGHLANDS IN VICINITY OF BEA

Relation of topography to major geologic fea

'WEST POINT AND VICINITY' SHEET, CORPS OF ENGINEERS, U.S. ARMY, 1:25,000, AMS 4, 1944.

OF BEAR MOUNTAIN, NEW YORK

major geologic features

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 61, PP. 137-190, 14 FIG., 5 PLS. MARCH 1950

STORM KING GRANITE AT BEAR MOUNTAIN, NEW YORK

BY KURT EMIL LOWE

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Abstract

The Storm King granite at Bear Mountain, New York, occupies the core of a syncline in the earlier crystalline complex of the Hudson Highlands. It is part of a larger synclinal pluton whose accordance with the northeast-plunging structure is shown by conformable contacts and mineral alignment in both the granite and the country rocks.

Reasonably constant mineral composition, flow structures of early hornblende crystals, and inclusions confined to the margins of the granite body indicate magmatic intrusion. Absence of secondary foliation and lack of tectonic fabric patterns in the granite suggest post-tectonic emplacement.

The concept of "exchange of space" between the rising magma and the country rocks subsiding into the emptying magmatic chamber is proposed to account for the lack of evidence indicating lifting of the overlying rocks by forcible injection of the granite.

In contrast to most fusion tectonites described, quartz *c* axes do not have clearly preferred orientations. Anomalous relations between dimensional and lattice orientation of the larger quartz grains are tentatively explained by differential solution of quartz by hydrothermal end-stage products of magmatic (Storm King granite) origin.

Field and statistical studies of principal joint systems reveal concentrations of longitudinal, cross, and diagonal fractures. Joint patterns in the Storm King granite simulate those in the Highlands complex. Several possible mechanisms are discussed to explain their development.

Interpretation of faults in conjunction with other structural data is used to distinguish between the effects of Precambrian and Paleozoic deformation intervals.

Introduction

Geographical Setting

Bear Mountain rises, only 40 miles north of New York, directly from the west shore of the Hudson River estuary to 1305 feet elevation. Its dome shape (Pl. 2), accentuated by a bare rock crest and precipitous flanks, makes it a prominent summit of the Hudson Highlands. This geographical term has been applied rather loosely to the New York State portion of the crystalline "Highlands", a chain of rugged mountains extending about 140 miles from Reading, Pennsylvania, through northern New Jersey and southeastern New York into western Connecticut (Fig. 1). Geomorphically the Highlands are known as the Reading Prong of the New England Upland.

Bear Mountain and its immediate vicinity are located in the Palisades Interstate Park, a forest preserve, while the Highlands region north and northwest of Queensboro Brook and lower Popolopen Creek (Hell Hole), excluding Fort Montgomery and other villages along the Hudson River, is part of the Reservation of the U. S. Military Academy at West Point (Pl. 1).

The Perkins Memorial Drive leading to the top of Bear Mountain is the most recent addition to the rather dense network of roads and deserves special mention. It is a one-way road spiralling about the mountain and affords magnificent views and excellent rock exposures along its entire length of 5 miles. The ascending and descending portions of the drive will be referred to in this paper as (up) and (down) respectively (Pl. 5).

Previous Work

William W. Mather (1843) first described igneous and metamorphic rocks in the Hudson Highlands and gave many specific locations of their occurrence. Scattered observations in the reconnaissance survey of Rockland County (Fig. 1) by Ries (1895) added to the knowledge of the Hudson Highlands crystallines. Kemp (1888a; 1888b; 1911; 1912) supplied specific data on the petrography and structure of these rocks. Stewarts's (1908) studies of the magnetite deposits east of the Hudson River gave some valuable information on a local scale. Berkey and Rice (1919), presenting the first

detailed and comprehensive picture of Highlands geology, undertook a systematic treatment of the complex rocks, outlined the sequence of geologic events, and correlated the

Significance of the Area

The Bear Mountain area was singled out for investigation after the writer had completed an

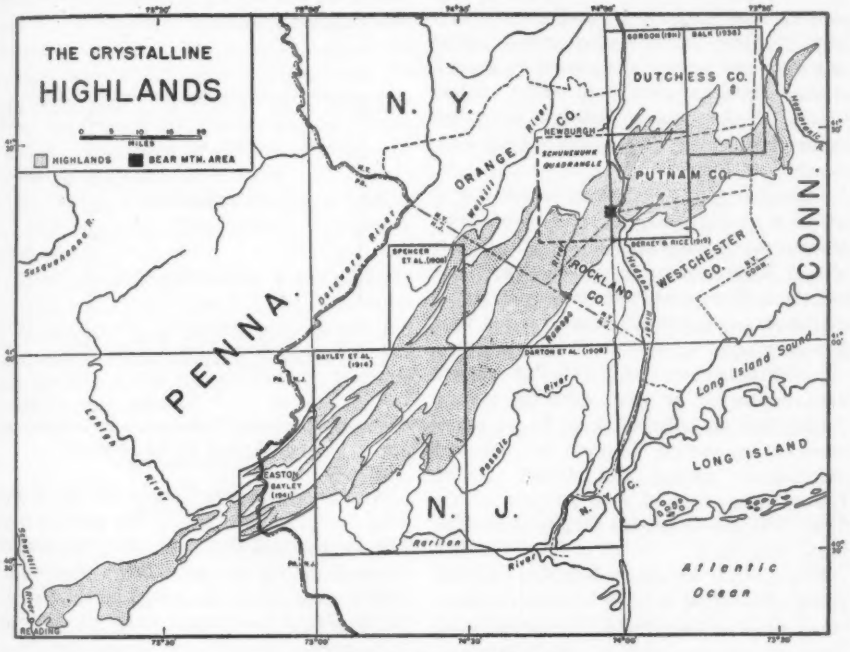

FIGURE 1.—INDEX MAP

rocks with those of contiguous and more distant localities of similar geologic history. The eastern portion of the writer's area of investigation appears on the geologic map of Berkey and Rice but had evidently not been studied in detail.

Colony (1921) largely supported Berkey and Rice and added many details of structure. Between 1925 and 1935 Professor Colony directed a program of geological field studies in the Schunemunk quadrangle (Fig. 1) in both the crystalline rocks of the Highlands and the broad belt of sedimentary formations to the northwest; the results of this study are not known.

Thompson's recent geomorphic study of the Hudson River gorge (1936) contained many structural data of particular value and pertinence to the writer's problem.

extended reconnaissance survey of the Highlands in the Schunemunk quadrangle between the Hudson and Ramapo rivers (Fig. 1). The fundamental concepts of Highlands geology established by Berkey and Rice were found generally applicable, but a detailed study was needed of the relations among certain principal crystalline formations with particular emphasis upon their relatively little known structure. The Storm King granite was selected because of its uniform structural and lithologic character and its topographic prominence. Extensive road construction at Bear Mountain provided unusually fine exposures of this granite including outcrops of critical intrusive contacts. Furthermore, this region proved potentially most significant because the folded structure of the Hudson Highlands crystallines could be demonstrated.

Field and Laboratory Work

General.—Field data were plotted on the very accurate topographic map of the Palisades Interstate Park, Harriman and Bear Mountain sections, New York, U. S. Geological Survey, 1928, 1:12,000, 10-foot contour interval. Bear Mountain and vicinity are covered by sheet 2 of this provisional (black and white) edition. Clarity of reproduction, however, demanded the use of a different base for the geologic map. (*See* Map co-ordinates.)

The writer spent 120 days in the field over a period of several years, but most of the investigation was accomplished during the summer of 1942. Regional reconnaissance took up about two thirds of the time, while the remainder involved the more detailed survey of the Bear Mountain area.

In the laboratory nearly 200 thin sections were examined; 20 were oriented sections studied with the universal stage. In the preparation of petrofabric diagrams an improved counting-out method was devised (Lowe, 1946c). Four chemical rock analyses of the Storm King granite and the Canada Hill granite phase were made.

Evaluation of structural field data required construction of 14 cellulose acetate and cardboard models (Lowe, 1946a). A new statistical approach was employed to obtain more accurate results on the mean orientation of linear structure (Lowe, 1946b).

Stereoscopic study of aerial photographs proved helpful in tracing the larger structural features of the region (Pl. 1).

Map co-ordinates.—The topographic base maps used in this paper are taken from the West Point and Vicinity, N. Y., sheet of the War Department, Corps of Engineers, U. S. Army series, 1:25,000, 3rd edition (AMS 4), 1944. Allowance should be made for slight distortion due to photographic reproduction.

Co-ordinate lines refer to the 1000-yard military grid, U. S. zone A, in accordance with the U. S. Coast and Geodetic Survey, Special Paper 59. The last three digits of the grid numbers are omitted. The geographical co-ordinates are indicated by ticks on the margins and cross markers in the body of the map (Pl. 5).

Exact locations of points referred to in the text are given by military grid co-ordinates to the nearest 10-yard interval (Heavey, 1941, p. 12). The *X* or east-west co-ordinate always precedes the *Y* or north-south co-ordinate, and all but the last two digits of the 1000-yard grid numbers are dropped.

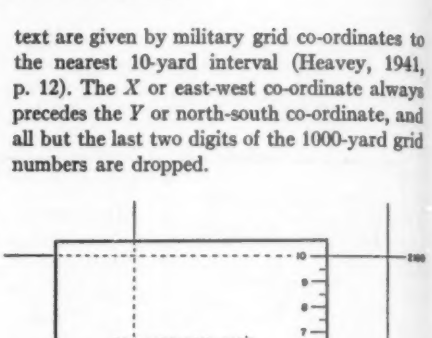

FIGURE 2.—READING CO-ORDINATES WITH GRID CARD
Co-ordinates of T: (07.77–99.08)

An auxiliary grid card is provided (Pl. 4) and may be cut out for use on the geologic map (Pl. 5). Figure 2 illustrates the procedure for determining the location of the observation tower on top of Bear Mountain, namely, (07.77–99.08). The *X* co-ordinate is read at the lower left corner (*A*) of the grid square, and the *Y* co-ordinate at point T on the map. Since the grid card is graduated in steps of 50 yards, the 10-yard intervals must be found by sight interpolation.

Acknowledgments

Prof. C. P. Berkey suggested the study of the crystalline rocks of the Schunemunk quadrangle to the writer, but the completion of the present problem would not have been possible without the guidance and inspiration of Prof. S. J. Shand and Prof. W. H. Bucher.

Thanks are extended to Mr. A. K. Morgan, General Manager of the Palisades Interstate Park Commission, and his staff for permission to pursue field work in the Harriman and Bear Mountain sections of the Park and for many other courtesies.

The co-operation of the U. S. Military Academy at West Point in encouraging geological

investigations on the Military Reservation despite difficult wartime conditions is most gratefully acknowledged. The writer wishes to express his appreciation to Col. L. E. Schick, Lt. Col. W. J. Regan, and Lt. Col. M. S. Dickson of the Department of Military Topography and Graphics of the Academy for furnishing the newest maps of the region and permitting the writer to study pertinent aerial photographs from the Department files. Permission to publish the photograph of Plate 2 was also obtained through their good offices.

The writer is indebted to Mr. Julius Kaikow and Mr. Herbert D. Glass for valuable assitance in the field.

Three chemical analyses (A, B and C, Table 1) were made by Dr. Lee C. Peck, Rock Analysis Laboratory, Department of Geology and Mineralogy, University of Minnesota. Mr. George Rév of Columbia University prepared excellent thin sections.

Geomorphology

Subaerial Erosion

In view of the long, continuous exposure of the Highlands to subaerial erosion (perhaps since early Mesozoic times) present topographic features exhibit the effects of structural and lithologic control to a high degree. In some instances even minor internal rock structures (*e.g.*, obscure platy flow structures in the Storm King granite) are emphasized by the development of subsequent erosional surfaces. Thus geomorphic expression is frequently used to interpret concealed or questionable subsurface structures.

It was thought advantageous, therefore, to discuss the several significant landforms in conjunction with the controlling factors of structure and lithology.

Effects of Glaciation

Evidence of Pleistocene continental glaciation is abundant in the Hudson Highlands (Berkey and Rice, 1919, p. 145; Thompson, 1936; p. 1844; Denny, 1938).

Upland surfaces such as Bear Mountain and The Torne show glacial polishing. Glacial striae, however, are not nearly so common as one would expect, because weathering and exfoliation on the exposed granite ledges have effectively destroyed such markings. Striae strike S. 30° E. as Thompson (1936, p. 1844) has shown. Chatter marks are prominent on nearly horizontal granite surfaces on top of Bear Mountain. The cirquelike basin of Brooks Lake northwest of Fort Montgomery (Pl. 1) is considered due to glacial plucking. This bedrock depression also may have been a plunge basin for glacial meltwaters during the later stages of its development. Hessian Lake likewise was deepened by the scouring action of the ice sheet.

Glacial boulders are scattered over the entire area. Till and drift have accumulated to considerable depth in the Doodletown Brook valley and in the Queensboro Brook-Popolopen Creek valley east of Weyants Pond. A glacial stream delta is being worked as a gravel pit at the base of The Torne east of the confluence of Cranberry Brook and Popolopen Creek (Pl. 1).

Weathering

A characteristic weathering phenomenon of the Storm King granite is its prominent exfoliation. Sheeting is parallel to the existing topographic surfaces, and individual spalled layers may be several feet thick (Pl. 3, fig. 1). The process of "unloading" seems best to explain the development of sheeting in the Storm King granite.

Chemical agents also play an important part in the weathering of this granite. Solution pitting on level glaciated surfaces indicates their effectiveness since deglaciation. Where the ice failed to remove the residual granite mantle the soil is coarse and "rubbly". The individual weathered fragments are elongated and preserve the internal alignment of the dark minerals in the fresh rock. Petrographic examination indicates that the crumbling of the rock fabric is due to kaolinization of the feldspars rather than alteration of hornblende or biotite. There is also some evidence that "rubbly" weathering is more pronounced in the syenitic facies of the Storm King granite and is associated with coarse rock texture and prominent development of linear and platy flow structures. Spheroidal weathering is sometimes caused by this chemical attack.

Deeply weathered zones in the Storm King granite grading rapidly into relatively fresh rock on all sides have not been explained satisfactorily. It has been suggested that they represent remnants of preglacial, residual soil which somehow escaped removal by the ice. But similar, though narrow, zones were observed along recently exposed sheeting surfaces which evidently were channels for ground-water percolation. Thoroughly decomposed Storm King granite is exposed at the base of a 30-foot cliff of relatively less weathered Highlands complex rocks (Canada Hill granite phase) into which it was intruded (at 09.60–99.37). Some evidence, therefore, suggests that such rotten zones may be produced by subsurface agents.

Huge talus boulders, some 20 feet or more on a side, are common along the roads skirting the base and flanks of Bear Mountain. They are generally joint blocks which, loosened by processes of mechanical weathering, tumbled from the steep rock scarps above. Some of these rock falls must have had landslide proportions as suggested by the ridgelike pile of large talus debris at the southwest end of Hessian Lake (09.12–99.24).

Petrography and Petrology

Hudson Highlands Complex

General statement.—The term "Hudson Highlands complex" or "Highlands complex" includes the entire sequence of crystalline rocks older than the Storm King granite. Colony (1921, p. 37) considered the Highlands complex synonymous with Highlands crystallines—*i.e.*, including the Storm King granite. At Bear Mountain and probably elsewhere the complex behaved like a single structural unit which guided and controlled the emplacement of the later Storm King granite. This gneissoid granite was intruded conformably with the structure of the "country rocks" of the Highlands complex. We are, therefore, less concerned with the successive events which gave rise to the various lithologic units of pre-Storm King age than with the structure of these rocks at the time of the Storm King granite invasion.

Grenville metasediments.—These quartzitic, micaceous, and calcareous beds comprise the oldest formation in the Hudson Highlands. Grenville age of these rocks is assumed (Berkey and Rice, 1919, p. 105, 140).

The rocks are characteristically layered (Pl. 3) and have been intensely metamorphosed during several periods of regional deformation and igneous invasions. A great variety of rock types resulted. Berkey and Rice (1919, p. 49–51) described only a few of the more characteristic types. At Bear Mountain where the Grenville series constitutes the major part of the Highlands complex, biotite, hornblende, epidote, and graphite schists and garnetiferous biotite gneisses are common.

One of the most unusual metamorphics at Bear Mountain is a coarsely crystalline eclogite[1] occurring as an intercalated layer, 4 to 8 feet thick, in the Grenville series. Dark-green pyroxene, ruby-red garnet, and metallic-gray graphite give this rock a striking color and appearance.

This eclogite layer is exposed on the east side of US 6 about 900 feet south of the Perkins Memorial Drive entrance (at 06.96–99.68). It also appears, evidently as part of the same layer, on both sides of the Perkins Memorial Drive (up) at (07.20–98.83). Characteristically rounded garnet crystals seem to be roughly aligned parallel to the "bedding." The large poikilitic pyroxene crystals, however, show no comparable mineral parallelism or alignment.

It is mainly composed of diallage (ferroan diopside) and garnet with subordinate graphite, magnetite, and quartz. The large euhedral to subhedral pyroxene grains are poikilitic and contain garnet crystals, graphite flakes, quartz grains, and altered hornblende remnants. Garnet likewise is host to the other minerals. Graphite flakes, with few exceptions, occupy cleavages and fractures or follow crystal boundaries. The microscopic texture is interlocking to uneven granular.

In many localities (US 9W opposite Iona Island, Doodletown Brook gorge, etc; Pl. 5) intercalated lenticular beds of graphitic marble showing intense plastic flow deformation are exposed. The calcareous and foliate facies of the

[1] The term "eclogite", as applied to this curious rock is used advisedly, since the dominant pyroxene appears to be diallage rather than omphacite. On the other hand, this name was also given to a diopside-garnet rock from Norway (Johannsen, 1938, IV., p. 462).

Grenville series behaved as incompetent layers (Fig. 5) compared with the rocks more clearly associated with igneous activity (mainly granites, granite gneisses, and pegmatites). Hence these beds appear most frequently distorted, crumpled, and drag-folded. They are also most readily affected by chemical weathering and are therefore differentially eroded to produce most of the topographic lows of the region (Fig. 6).

Pochuck diorite phase.—Berkey and Rice (1919, p. 51, 57) showed that the oldest igneous representative distinguishable in the Hudson Highlands is essentially a diorite, but they indicated that it is practically always intimately associated with Grenville metamorphics. Hornblende-plagioclase gneiss layers seen at Bear Mountain are believed to represent Grenville rocks reworked selectively by this diorite invasion. They correspond to Colony's "Pochuck-Grenville" (1921, p. 50).

The presence of a diorite parent rock of unquestionable magmatic origin could not be demonstrated in the Bear Mountain region. The writer, therefore, prefers to use the term "phase" for this earliest igneous interval. There is no intention, however, to question the validity of the intrusive concept. Colony (1921, p. 49–53) described magmatic diorites in the magnetite mines of the Highlands. He also explained associated pyroxenites, hornblendites, syenites, and even an unusual type of granite as differentiation products of the Pochuck magma.

The writer found a typical hornblende-pyroxene diorite with oligoclase-andesine feldspars on the west shore of Lake Tiorati (Fig. 3). It is a fresh, coarse-grained, green-gray rock of uniform composition and structure, showing linear flow alignment of the subhedral to euhedral dark minerals. This rock appears to be older than the surrounding granite gneiss (probably Canada Hill phase) which sharply truncates the diorite and gives the impression of having been plastically squeezed around the earlier massive intrusive.

The Highlands complex is cut by a variety of pegmatites, and it is not always easy to relate individual ones to a specific magmatic interval. The writer, however, believes that a coarse, curiously mottled, light- to medium-bluish-gray, highly feldspathic pegmatite can be related to the Pochuck diorite phase. It is mainly oligoclase-andesine with variable amounts of quartz. Graphite flakes (up to ½ inch diameter) in many places occupy feldspar cleavages and quartz fractures or wrap around individual mineral grains.

Canada Hill granite phase.—In place of the Canada Hill, Reservoir, and Mahopac granites described by Berkey and Rice (1919, p. 52–56), the writer includes under the term Canada Hill granite phase all rocks representative of granitic igneous activity in the Hudson Highlands after the Pochuck diorite phase and earlier than the Storm King granite intrusion. The writer believes such broad generalization is necessary because of the difficulty of distinguishing among the three very similar granite types in the field. Berkey and Rice considered the Mahopac granite a facies of the Reservoir type, and Colony (1933, p. 25) questioned the existence of the Reservoir and Canada Hill as two separate granites.

Work in the Highlands west of the Hudson River has convinced the writer that only the Canada Hill type granite can be recognized with certainty and that the Reservoir and Mahopac granites are not separable from it. Furthermore there is evidence that at least some of the rocks called Canada Hill are perhaps the result of granitization[2] of certain Grenville sediments. The term "granite phase" is therefore preferable, because it includes all granites derived from the same source whether formed from a true magmatic melt or by granitization of pre-existing rocks.

Perhaps the most typical representative of the Canada Hill phase is a medium-gray, medium-grained biotite granite. The white and gray feldspars are principally albite-oligoclase and perthite with orthoclase and microcline sometimes present in appreciable amounts. Gray quartz is an essential constituent, and violet-red to dull-red garnet is an abundant accessory. Biotite flakes are characteristically oriented in layers which give the rock a faint

[2] "Granitization" (including the adjective "granitized") is used in the sense of Grout's (1941, p. 1540) broad definition: "... granitization includes a group of processes by which a solid rock (without enough liquidity at any time to make it mobile or rheomorphic) is made more like granite than it was before, in minerals, or in texture and structure, or in both."

to excellent foliation structure depending on the quantity of biotite. Weathered surfaces are dull gray and tend to exfoliate less than the Storm King granite does.

The most common pegmatitic facies of the Canada Hill phase is a coarse-grained, white to light-gray, quartz-feldspar (acid plagioclase and perthite) rock which resembles alaskite and carries patches of red-brown garnets. Quantities of epidote, zoisite, tourmaline, and biotite may be present.

The Canada Hill granite (Berkey and Rice, 1919) was derived from a very fluid and chemically most active magmatic melt of granitic composition which became "contaminated" by "magmatic absorption or syntexis" of the invaded rocks (p. 37–40). Granitization (evidently used in the sense of metasomatism) of the country rocks by "mineralizers or emanations in a very fluid condition, at high temperature and under great pressure" is also mentioned (p. 41). On the whole, however, the impression is given that the magmatic stage was always involved and that the "granitized" rocks can generally be distinguished from the "true" granite.

The writer suggests that metasomatism[3] could have been responsible for the formation of rocks which appear identical with the Canada Hill granite of magmatic origin.

Plate 3 shows modified Grenville sediments along Perkins Memorial Drive (up) at Bear Mountain. Layers of granitic composition alternate with biotite schists and biotite-hornblende gneisses. The layers of different composition are sharply defined, have uniform thickness, and can be traced for more than 600 feet at this locality. The granitic layers are of particular interest. Some (Pl. 3, fig. 5) are remarkably uniform, medium-grained quartz-feldspar rocks containing discontinuous stringers of slightly coarser pegmatitic material of the same composition invariably oriented parallel to the rock structure. If such rocks were encountered where their intrusive origin is unquestionable, they would be called alaskites. Magmatic emplacement, however, does not seem to explain adequately the origin of these pseudo-alaskites, as the writer prefers to call them. Attention is directed particularly to the uniformly thin, continuous layers of well-foliated and jointed biotite schist and hornblende gneiss separating the relatively thick granitic layers (Pl. 3, fig. 5). The most careful scrutiny revealed no dikes or offshoots of the igneous-looking rocks into the adjoining fissile metamorphics. While the presence of such offshoots would not prove the mode of origin of the granitic layers their absence can be more readily accounted for by some process of selective reworking of sediments in place than by concordant magmatic intrusion at depth.

Under the microscope some significant departures from the subhedral to anhedral granular texture of the typical alaskite (Johannsen, 1932, II, p. 107) can be observed. Intergrown aggregates of feldspar (microperthite, acid plagioclase, and microcline) seem to have replaced the larger quartz grains producing deeply embayed outlines and veinlike structures along visible fractures. The quartz is clear and has uniform extinction. Also most of the larger feldspar grains not confined to a particular variety contain a profusion of unoriented inclusions of all the other types of feldspar.

These features suggest replacement and recrystallization of a pre-existing rock (possibly a rather pure arkosic sandstone) by hydrothermal emanations, rich in alkalies, from a magmatic source of perhaps granitic composition.

Biotite appears in some of the pseudo-alaskite layers (upper light-colored layer, Pl. 3, fig. 5) in sufficient quantities to make them megascopically indistinguishable from an intrusive biotite granite such as the typical Canada Hill granite of Berkey and Rice. Furthermore, the microscopic evidence of replacement action is not always present.

Storm King Granite

Normal type.—The Storm King granite represents the last major invasion of magmatic origin in the Hudson Highlands. In contrast to the great variety of rock types in the Highlands complex and their involved field relations, this granite occurs in large masses of rather uniform character. Hence, it is the most distinct litho-

[3] Metasomatism as defined by Lindgren (1925, p. 247) is "an essentially simultaneous, molecular process of solution and deposition by which, in the presence of a fluid phase, one mineral is changed to another of differing chemical composition".

logic unit encountered in the crystallines of the Hudson Highlands.

The typical Storm King granite (Berkey and Rice, 1919, p. 56) is a medium- to coarse-grained rock which is dull gray on fresh exposures, sometimes with a greenish to pinkish-buff tinge and a somewhat greasy luster. Its characteristically streaky appearance, the result of linear alignment of the dark minerals, constitutes one of the most constant structural criteria for the recognition of this granite in the field.

More than 60 per cent of the rock (by volume) is gray and reddish feldspars. Microcline, microcline-microperthite, and perthite predominate. Orthoclase and albite-oligoclase occur in relatively minor amounts. The abundance of potash feldspars is perhaps the most characteristic petrographic feature. Where microcline and microcline-microperthite are important constituents of the Highlands complex the influence of a near-by Storm King granite intrusion can nearly always be demonstrated. The quartz content ranges from practically none to about 30 per cent by volume.

Under the microscope most feldspar and quartz grains are anhedral. Subhedral to euhedral hornblende is the characteristic dark mineral, but augite and biotite may be present. Strong pleochroism of hornblende and biotite is typical. Common accessory minerals are zircon, apatite, and magnetite. Allanite was found in some thin sections.

The pegmatitic facies is very coarse-grained and distinctly pink to reddish; its minerals have no preferred orientation. Feldspars are of the usual Storm King varieties, and quartz is abundant. The hornblende content is lower than in the normal granite, but individual crystals may reach 12 inches in length. Graphic intergrowth of quartz and feldspar is not uncommon.

Weathered surfaces of the Storm King granite are buff to reddish. Long exposure to atmospheric agents produces marked kaolinization of the feldspars particularly in the quartz-poor facies. Despite this tendency toward decomposition, the Storm King granite is the most resistant crystalline rock in the Highlands.

The distribution of the syenitic facies of the Storm King granite appears to be haphazard. The quartz content decreases slowly so that the boundaries between the granitic and syenitic portions of the rock are usually imperceptible. Some irregular masses of pegmatite within the Storm King granite are surrounded by relatively narrow zones (rarely more than 1 foot wide) of dark syenite, uncommonly rich in hornblende, augite, and biotite. Yet the only reasonably constant relationship seems to be the association of high quartz content with the red feldspars (mostly microcline and microcline-microperthite) as observed by Berkey and Rice.

Contact facies.—Certain mineralogical changes in the Storm King granite near contacts with the Highlands complex are sufficiently distinct and constant to merit special mention. The most common features of this contact facies are:

(1) Dark minerals decrease in quantity and in some places disappear
(2) Biotite (slightly to moderately pleochroic) appears instead of hornblende and augite as the principal dark constituent
(3) Potash feldspars (microcline and microcline-microperthite) are relatively less abundant than more sodic varieties (perthite and albite-oligoclase)
(4) Quartz content is often high
(5) Garnet and in some places graphite are important accessories
(6) In some instances intense chloritization along quartz fractures and feldspar cleavages imparts a dark-green color to the rock. Brownish-green quartz is always present and has a decidedly greasy luster. Biotite partly altered to chlorite is the dark mineral. The best example of this rather unusual type of contact facies may be observed on Perkins Memorial Drive (up) between (07.12–99.16) and (07.20–99.22). Here chloritization appears to have been most intense in a zone about 150 feet wide and projecting perhaps 200 feet into the granite roughly perpendicular to the inferred contact.

Most of these contact features are undoubtedly the result of reaction between the Storm King magma and its wall rocks. Cross assimilation (reactive solution and precipitation) involving an exchange of certain components (Shand, 1943, p. 95) would account for the field evidence. Type of contact facies and composition of the adjacent country rocks seem to be

TABLE 1.—CHEMICAL COMPOSITION AND NORMS OF HUDSON HIGHLANDS GRANITES

	Canada Hill granite	Storm King granite		
	A	B	C	D
		contact facies	normal type	normal type
	L. C. Peck, analyst			K. E. L., analyst
SiO_2	72.39	72.35	73.35	73.45
Al_2O_3	15.48	14.04	13.07	12.72
Fe_2O_3	.17	.57	.74	1.41
FeO	1.66 (1.83)	2.22 (2.79)	2.29 (3.03)	2.12 (3.53)
MgO	.42	.27	.08	.02
CaO	1.96 (2.38)	1.29 (1.56)	1.11 (1.19)	1.32 (1.34)
Na_2O	5.02	4.03	2.98	3.67
K_2O	2.20 (7.22)	4.44 (8.47)	5.62 (8.60)	5.16 (8.83)
H_2O	.33	.24	.23	.26
TiO_2	.24	.27	.20	.18
P_2O_5	.08	.07	.06	not det.
MnO	.05	.03	.03	not det.
	100.00	99.82	99.76	100.37

NORMS

Normative Minerals	A	B	C	D
quartz	29.16	27.72	30.70	29.10
orthoclase	12.79	26.13	33.25	30.10
albite	42.44	34.06	25.15	30.92
anorthite	9.17	5.84	5.14	3.06
corundum	1.53	.51	.18	—
apatite	.20	.17	.13	—
diopside	—	—	—	3.29
hypersthene	3.75	3.87	3.51	.81
magnetite	.23	.93	1.07	2.07
ilmenite	.46	.46	.38	.30
water	.33	.24	.23	.26
	100.05	99.93	99.74	100.41
Norm	I.4.2.4. Lassenose	I.4.2.3. Toscanose	I.4.2.3. Toscanose	I.4.2.3. Toscanose

A: Canada Hill granite, 540 feet elevation, US 6 (06.95–99.68), approximately 350 feet below Storm King contact.

B: Storm King granite, contact facies, 700 feet elevation, Perkins Memorial Drive (up), (07.07–99.28), approximately 40 feet above lower contact.

C: Storm King granite, normal type, 930 feet elevation, Perkins Memorial Drive (up), (07.18–99.31), approximately 400 feet above lower contact.

D: Storm King granite, normal type, 460 feet elevation, US 6 (07.57–00.34), approximately 250 feet above lower contact.

related. Where the Storm King granite intruded the more basic Grenville schists and gneisses, the near-contact portions of the intrusion appear to be particularly rich in biotite, garnet, graphite, and plagioclase feldspars. (*See* (2), (3), and (5) above.) Contact rocks containing few, if any, dark constituents, more potash-soda feldspars (perthite), and much quartz are

usually associated with the Canada Hill phase of the Highlands complex. (*See* (1), (3), and (4) above.)

The writer believes that local contact zones rich in chlorite and quartz can be explained more satisfactorily by the action of volatile end-stage products, escaping from the Storm King magma, upon the chilled borders of the intrusion.

Chemical analyses.—The literature contains no reference to the chemistry of the Hudson Highlands crystallines. Therefore, analyses were made of the Canada Hill and the Storm King granites (Table 1).

The comparative chemical relations between the Storm King granite and the Canada Hill granite phase of the Highlands complex are in harmony with the petrographic evidence. From the normal rock of the Bear Mountain intrusive across the lower contact into the most characteristic granite phase of the country rocks, quantitites of the most variable constituents change consistently. For instance, an increase of Na_2O is accompanied by a decrease of K_2O, while total alkalies drop only slightly. This accounts for the mineralogical change from microcline and microcline-microperthite in the normal Storm King granite to perthite with some microcline and plagioclase near the contact to predominant plagioclase and perthite in the Canada Hill phase. Higher CaO in the Canada Hill phase is at least partially due to a greater percentage of anorthite molecules in the plagioclase feldspars. The remainder is probably contained in garnet, an important accessory of the Canada Hill granite, which is also responsible for the relatively higher quantity of Al_2O_3 in this rock.

The appreciable difference in total iron oxide (FeO plus Fe_2O_3) content between the earlier and later granite appears to involve the presence of magnetite as a common accessory mineral in the Storm King granite, but not in the typical Canada Hill phase, and the distribution of the dark minerals in the two granites. Hornblende (by volume) is more abundant in the Storm King granite than biotite is in the Canada Hill phase.

Perhaps the most significant fact brought out by the chemical data is that the composition of B is intermediate, in almost all particulars, between that of A and C (Table 1). Thus the chemical evidence supports field and petrographic relations which indicate that the Storm King contact facies was produced by reaction between the intrusive and its wall rocks.

Although the two normal Storm King granite samples (C and D) were mineralogically rather similar, somewhat different values for Fe_2O_3, CaO, and Na_2O showed up in the chemical analyses. This probably means that the samples used for analysis were not large enough to be representative of such a coarse-grained rock. However, the presence of uncommonly large crystals of apatite and a good deal of magnetite in specimen D does account for the relatively high values of CaO and Fe_2O_3 obtained from this sample.

End-stage products.—Primary fractures in the Storm King granite are characteristically filled with minerals evidently derived from the same magmatic source.

Fine-grained aplite dikes composed of equigranular quartz, microcline, and microcline-microperthite with occasional small hornblende grains are common. This mineral assemblage definitely establishes the Storm King magma as the source of this rock.

Stringers of gray, coarsely crystalline quartz seem to be more numerous near contacts and tend to occupy cooling fractures parallel to the contact surfaces.

Epidote is one of the most common end-stage products. It may follow principal joint systems or penetrate the granite in all directions along even the finest fractures. Epidote serves as a most efficient healing agent in the more closely fractured portions of the granite. A common association of this mineral with the pink (quartz-rich) Storm King facies was also observed.

Chlorite and sericite coat fracture walls. They are readily slickensided and are useful aids in the recognition of post-crystallization movement along primary joints.

A most curious rock of presumably hydrothermal magmatic origin was discovered on the Seven Lakes Drive between (09.32–98.35) and (09.35–98.57). The dark-green, dense material, resembling basalt, has completely healed a brecciated zone of Storm King granite. It not only filled the spaces between the crush frag-

ments as well as minute fractures within them, but also penetrated the rock fabric along grain boundaries. Near the north end of the exposure a near-vertical dike, 2–3 feet wide, composed of the same material appears to branch out into the sheared granite and evidently represents one of the feeding channels.

The unusual mineral assemblage includes quartz, feldspars (orthoclase, albite to andesine, microperthite), amphibile, pyroxene, biotite, sericite, chlorite, epidote, zoisite, fluorite, carbonate, serpentine; and many other minor constituents. Beautiful hematite rosettes were observed on joint surfaces within the dike. The microscopic structure is fine granular to matted or felted. Flow alignment is prominent near the contacts and around inclusions. Epidote, zoisite, and carbonate are common in veinlets. Included fragments of Storm King granite have been intensely altered. Cloudy alteration zones in the feldspars are arranged concentrically, with relatively fresh remnants of microcline forming the centers of some. Fine-grained contact-metamorphic zones of granodiorite in the granite adjoining the dike were produced by this type of feldspar alteration producing oligoclase-andesine.

The writer cannot classify this rock specifically but does not doubt its hydrothermal magmatic origin. The question, however, remains whether it can be properly considered an end-stage product of the Storm King intrusion. The only intrusives of post-Storm King age in this region are dioritic basalt dikes which have neither the composition nor the insidious and reactive features of this rock. Thus, on negative evidence alone, Storm King origin must be assumed.

Dike Rocks

Basalt dikes cutting all the crystalline rocks are the only evidence of post-Storm King magmatic activity in the Hudson Highlands. They appear in large numbers throughout the Highlands terrane and show sufficient megascopic variations in composition to make a systematic petrologic study of these dikes desirable. The writer did not investigate them owing to lack of time and the relatively small number of such dikes encountered in the Bear Mountain area.

On the whole their composition varies from dioritic to camptonitic. Some more acid varieties were described by Kemp (1888a; 1888b; 1911; 1912).

Fresh exposures are dark greenish gray but soon weather buff. The rock varies from felsitic to finely crystalline. Cooling joints in intricate patterns are nearly always present, and shear zones are not uncommon. The dikes average 4–5 feet wide, but extremes of 2 inches (Perkins Memorial Drive (down), (08.16–98.58) and 50 feet (Hell Hole, 07.52–00.47) were observed.

The basalt evidently chilled rapidly and was clearly guided by open fractures in the crystallines. Contacts are always sharp, and evidence of contact effects is lacking, both in the basalt and in the wall rock.

Structural Geology

General Considerations

The most prominent feature of the Highlands structure is its persistent southwest-northeast trend which was recognized nearly 80 years ago by Cook (1868, p. 51). His diagram of ridge trends from Reading, Pennsylvania, to the Hudson River was correctly interpreted as topographic expression of the general rock structure. No profound structural break has been demonstrated in the complex crystalline rocks of presumably Precambrian age extending in a continuous chain from eastern Pennsylvania through northern New Jersey into southeastern New York (Fig. 1). Colony (1921, p. 7–30) documented the growth of this concept of fundamental structural continuity. Balk (1936) carried the structure northeastward from the West Point, New York, quadrangle (Berkey and Rice, 1919) across the Connecticut line to the Housatonic River.

The earliest structural data of any consequence seem to have been recorded by Kitchell (1855) in the Highlands of northern New Jersey. Cook (1868) presented this information together with many measurements of his own in the form of tables showing directions and angles of dip and pitch[4]. Cook (1883) and Britton (1886) gave average numerical values of strike, dip, and pitch[4] for that region, based on tabular

[4] "Pitch" is equivalent to "plunge" as used in this paper.

compilations. Much of this structural information was incidental to the detailed study of the Highlands magnetite deposits which represented an important economic factor in the early days of the Geological Survey of New Jersey.

Considering the early recognition of the importance of factual structure data by the New Jersey geologists, it is surprising that relatively few measurements of this type are available for the Highlands of the Hudson. Ries (1895) recorded some scattered structure measurements of the crystalline rocks of Orange County, New York, west of the Hudson River. Studies of the magnetite deposits of southeastern New York on both sides of the Hudson River (Stewart, 1908; Colony, 1921) added some information concerning the rock structures associated with the ore bodies. The construction of the Catskill aqueduct of the City of New York spurred the geological investigation of the Hudson Highlands but contributed all too few actual measurements of the rock structures encountered. The first systematic study of joints and orientation of basic dikes along the Hudson gorge was undertaken by Thompson (1936) who also appears to have been the first to record the presence and attitude of linear structure in the Storm King granite. Balk (1936) published numerous structure measurements of the Highlands gneisses east of the Hudson River in Dutchess and Putnam Counties of New York.

Definitions

In the absence of an adequately standardized and generally accepted terminology of structural geology, definitions of certain terms repeatedly used in this paper are given.

The attitude in space of a line is defined by two angles:

Bearing—the projection of a line onto the horizontal plane, measured from the geographical north (Bucher, 1944, p. 195),

Plunge—the angle from the horizontal to the line, measured in a vertical plane (Billings, 1942, p. 44).

A line *within* a given plane is also defined by the

Pitch—the angle that the line makes with a horizontal line in that plane (Billings, 1942, p. 135).

Linear structure—parallel alignment of the long axes of mineral or rock grains or other linear components in a rock.

Linear components—in this case, the individual elongate mineral or rock grains responsible for the linear structure.

Lineation—a descriptive and nongenetic term for any kind of linear structure within or on a rock (Cloos, E., 1946, p. 1).

Linear structure involving parallelism of axes of mineral or rock grains *within* a rock should be differentiated from parallel alignment of the elongate traces of such linear components *on* an exposure surface. In a recent paper (Lowe, 1946b) the writer proposed to limit the concept of "lineation" to describe traces of linear structure on the surfaces of rock exposures. Since then the critical study of lineation by E. Cloos (1946) has appeared. To avoid further confusion of terminology, the writer has adopted the very general definition of "lineation" by Cloos. The term "linear traces" is suggested to express the very real difference between the attitude in space of elongated, essentially linear, three-dimensional components and the orientation of the long axes of their surface traces.

Linear traces—parallel alignment of the long axes of elongate traces of mineral or rock components on an exposure surface.

"Linear traces", although primarily intended to describe traces of linear structure, could also refer to traces of planar structures where the spatial arrangement of the mineral or rock components is not known. Cloos (1946), for instance, shows the sketch of a granite gneiss specimen from Callander, Ontario (Fig. 3, p. 15) possessing strong linear flow structure, but no planar foliation. In discussing this specimen (bottom, p. 15), however, the author states:

> "On all sides a strong *foliation* is visible; top and bottom are almost structureless." (Writer's italics).

It would not be difficult to visualize an outcrop of this rock in which the critical top and bottom surfaces are not visible. Thus all exposures would show strong surface lineations due to traces of linear structure, but not as a result of planar foliation, though this spatial relation of the mineral components could not be determined from the particular outcrop. Hence even tentative reference to this lineation as "folia-

tion" is incorrect, since it suggests the presence of a planar structure. On the other hand, use of the descriptive term "linear traces" would imply that the surface lineations represent traces of an internal mineral parallelism which could be linear, planar, or both.

Linear flow structure or flow lines—linear structure interpreted as the result of primary flow in igneous rocks.

Planar structure or foliation—arrangement of mineral or rock grains in parallel layers (planes or curved surfaces) with or without linear alignment of the individual components.

Platy flow structure or flow layers—foliation interpreted as the result of primary flow in igneous rocks.

Structures in the Hudson Highlands Complex

Remnant bedding.—Investigators in the Highlands area have agreed that the layered crystalline rocks of this region were originally sedimentary. There has been much controversy, however, concerning the relationship between the now-visible layers and the original, presumably stratified structure of the ancient sedimentary rocks. Berkey and Rice (1919, p. 49) assigned the metamorphic equivalents of these early sediments to the Grenville series and described them as a group of intercalated quartzitic, micaceous, and calcareous beds characteristically banded, crumpled, and distorted.

A reasonable corollary to this concept, then, would be that the present conformable attitude of both foliate and nonfoliate crystallines (*e.g.*, silicated limestones) of the Grenville series essentially preserves their sedimentary structural character. Layered mineral structures, therefore, could be interpreted as the "true" bedding of the original sediments. The writer favors this idea while conceding the difficulty of proving this contention in the face of the long and complex history of the region, which obviously was subjected to several extensive igneous invasions and deformational periods at depth. Plastic deformation, recrystallization, migmatization, and development of secondary foliation could well have destroyed all evidence of original bedding and even have produced apparent conformable relations where none existed. In fact, the very complexity of the processes involved has made it impossible to establish any adequate stratigraphic succession and correlation for the Grenville rocks of the Highlands.

Yet the Bear Mountain region has yielded some evidence to support the concept of "remnant bedding". The writer distinguishes between layered structures suggesting direct relation to original stratification and those exhibiting pronounced foliation which may be of secondary (deformational) origin. As applied to silicated limestones, for instance, "remnant bedding" implies primary structure, but not necessarily original attitude and thickness.

Plate 3 shows the rocks of the Highlands complex along the lower Perkins Memorial Drive (up) between (07.28–99.00) and (07.42–98.88) directly below the contact with the Storm King granite. The uniform dip as well as the sharp, though gently undulating margins of the individual layers, which can be traced continuously for about 600 feet along the strike, suggest a "bedded" appearance. Petrographic study supports the sedimentary origin of the rather massive light-colored layers (pesudoalaskite). Crude planar alignment of the quartz and feldspar crystals parallel to the "bedding" is believed to have been caused by recrystallization under deformational stress which formed the relatively broad synclinal structure of which these rocks are a part. Shearing appears to have been insignificant as indicated by the lack of the usual strain effects. Evidence of plastic flow is likewise absent. If this argument be correct, the conformable attitude of schistosity in intercalated micaceous layers, which might represent original pelitic beds, suggests an inherited "bedding".

A somewhat similar relationship is shown by

PLATE 2.—BEAR MOUNTAIN, NEW YORK

FIGURE 1.—BEAR MOUNTAIN AND SOUTHERN GATEWAY OF HUDSON RIVER GORGE
Low oblique air photo from 5000 feet altitude looking due east. Distance from Queensboro Lake to Anthonys Nose is 3½ miles. Courtesy U. S. Military Academy, West Point, New York.

FIGURE 2.—OUTLINE SKETCH OF FIGURE 1
Areal distribution of Storm King granite shown by stippling.

FIGURE 1

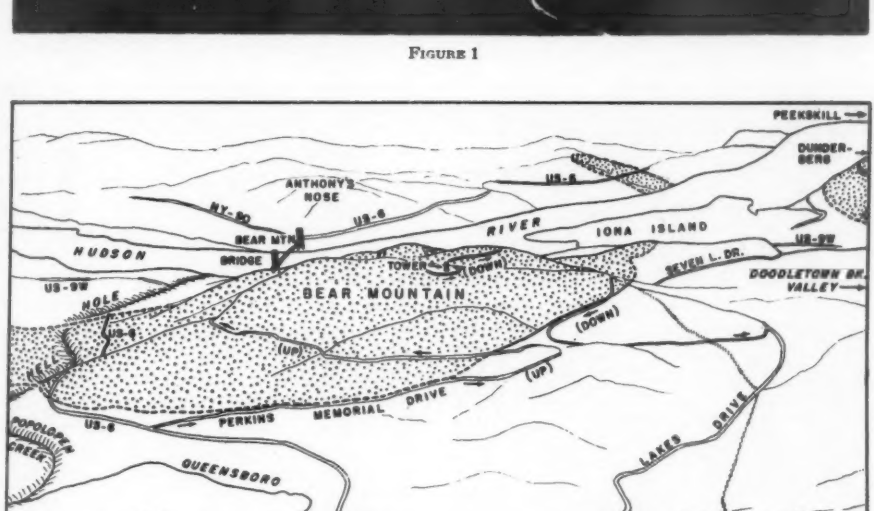

FIGURE 2

BEAR MOUNTAIN, NEW YORK

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

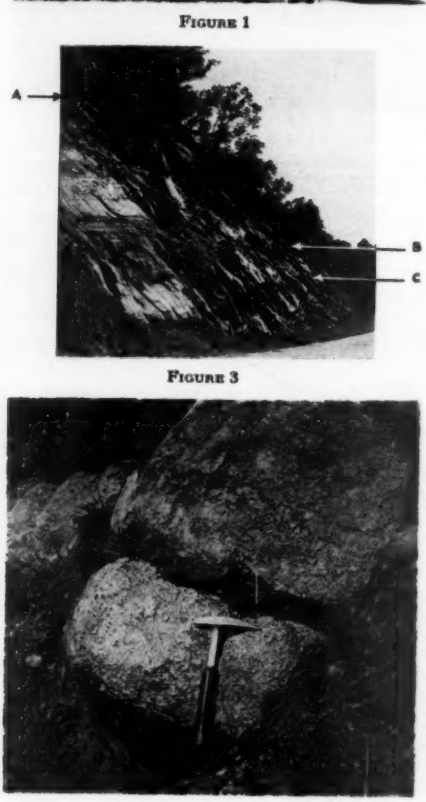

FIGURE 5

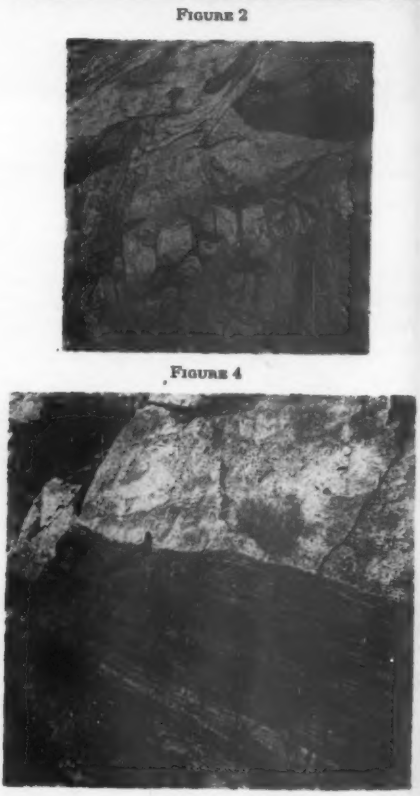

FIGURE 6

LOWER CONTACT OF STORM KING GRANITE WITH HIGHLANDS COMPLEX

several highly silicified crystalline limestone layers along US 6 between the entrance of the Perkins Memoral Drive and Queensboro traffic circle (06.78–99.33 and 06.31–98.96). In this case the sedimentary origin is established by lithology while the preservation of original bedding is suggested by the absence of plastic flow features, usually quite evident in the calcareous rocks of the Grenville series. Again the planar mineral structures in the adjacent crystalline foliates are conformable. This structural phenomenon may be considered characteristic of the Highlands complex in the Bear Mountain region. Therefore no distinction was made between remnant bedding and foliation in recording structural measurements (Pl. 5), particularly since the present investigation is primarily concerned with the later intrusive phase of the Storm King granite.

Foliation and folds.—The Highlands complex in the vicinity of Bear Mountain is a relatively broad syncline or synclinorium plunging about 40° NE. The mountain proper, which is largely Storm King granite filling the core of the syncline, occupies much of the area mapped in detail (Pl. 5). Therefore available structure measurements from the surrounding Highlands complex rocks do not give wholly satisfactory proof of the larger structure. Its synclinal attitude, however, is indicated by the gradual change in strike of foliation from N.-S. in the Weyants Pond-Queensboro Lake region to E.-W. and NE.-SW. in the lower Doodletown Brook valley. Dips are consistently northeast and appear to be gentlest near the synclinal axis east of Queensboro Lake.

The topographic trends of the Highlands ridges are closely related to the strike of the crystalline gneisses. (*Cf.* Cook, 1868, p. 51.) The parallel ridge crests in the northwest portion of Plate 1 reflect the general monoclinal structure of the rocks mentioned by Colony (1921, p. 32). To the southeast toward the Hudson River, however, a remarkable disturbance of this uniform regional trend can be observed.

Beginning with Turkey Hill west of Weyants Pond nearly continuous ridge crests represented by Summer Hill, West Mountain, the Timp, and Dunderberg follow an arcuate pattern about Bear Mountain as a center. This curving trend outlines the base of the Bear Mountain syncline which is exposed to a maximum width of about 5 miles along a line from the Hudson River at the base of Dunderberg to Stillwater Hill in the upper Popolopen Creek valley (Pl. 1).

The west shore of the Hudson River north of Fort Montgomery yields similar good evidence of this large plunging synclinal structure (Fig. 10). Not only does the synclinal structure become apparent, but its axial trend changes from NE. at Fort Montgomery to N.-NNE. at Highland Falls. This feature is believed to be related to movements along the Timp Pass-Hudson River fault.

The regional dominance of monoclinal structures with steep southeasterly dips was pointed

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PLATE 3.—LOWER CONTACT OF STORM KING GRANITE WITH HIGHLANDS COMPLEX

FIGURE 1.—LOWER CONTACT OF GRANITE

Contact surface, N.45°W.,40°NE., dips away from observer. Note prominent exfoliation of granite. Perkins Memorial Drive (up), 920 feet elevation, (07.28–99.00), looking north.

FIGURE 2.—CLOSE-UP OF CONTACT

Sharp, conformable contact dips 40° to left. Flow structures in granite parallel to contact and foliation in country rocks (lower right). Looking southeast from road.

FIGURE 3.—HIGHLANDS COMPLEX BELOW CONTACT

Granite contact at A (Fig. 2), Grenville metasediments below. "Bedded" appearance produced by crudely foliated granitic layers (light) alternating with biotite and hornblende schists and gneisses (dark). Looking southeast from road.

FIGURE 4.—MARGINAL INCLUSION IN GRANITE

Tabular inclusion of Canada Hill granite phase parallel to contact, dips 40° away from observer. Margins indicated by truncated longitudinal (dip) joints and ends of hammer. Later cross-cutting dip joint at left. Looking northeast from road, 8 feet above contact (Fig. 1).

FIGURE 5.—CLOSE-UP OF GRENVILLE METASEDIMENTS

Pseudo-alaskite (Canada Hill phase) with intercalated layers of hornblende gneiss (upper) and biotite schist (lower). Foliation conformable throughout. View at B (Fig. 3), looking north, (07.37–98.88), 40 feet above road and 28 feet (true) below contact.

FIGURE 6.—CLOSE-UP OF GRENVILLE METASEDIMENTS

Sharp contact between biotitic diorite gneiss (Pochuck phase) and biotite granite (Canada Hill phase). Foliation conformable in both. View at C (Fig. 3), looking north, 20 feet above road and 55 feet (true) below contact.

out by Colony (1921, p. 32) without ruling out the likelihood of folding when he stated:

> "There are exceptions to this general monoclinal habit which seem to indicate the probability of a very definite folded structure which is, however, somewhat obscure..."

This argument was largely based on minor, mostly synclinal folds reported from the subsurface study of iron mines in the New Jersey and Hudson Highlands. The writer has found no references in the literature concerning large folds similar to the Bear Mountain syncline which are clearly reflected in the topography. At least one other example may be cited from the writer's field reconnaissance at the south end of Lake Tiorati, 5.5 miles southwest of Queensboro traffic circle along the Seven Lakes Drive (Fig. 3). The axis of folding plunges 15° to 20° approximately N. 39° E. The nose of the syncline, particularly well exposed, exhibits characteristically steep outer and gentle inner slopes developed on highly granitized Grenville metasediments. Possibly large folds of this type are not nearly so rare as has been intimated. A careful geomorphic study of the region with the help of large-scale maps and aerial photographs would seem to be a method of approach worthy of consideration.

Another somewhat unusual feature of the Bear Mountain syncline is its relatively high angle of plunge (40°) compared with the average of 30° reported by Britton (1886) for the Highlands of northern New Jersey. The axial plunge of the Tiorati syncline (15°–20°) appears to be closer to the norm of 25° observed by the writer elsewhere in the Hudson Highlands. This apparent steepening of the tectonic axis may be the result of differential sagging of the structure during emplacement of the Storm King magma (*see* Origin of the Storm King granite), or it could be due to later tilting in response to thrust movements along the Timp Pass-Hudson River fault to the southeast and the Crown Ridge fault to the northwest (Fig. 10).

In the Fort Montgomery area between Brooks Lake and the Hudson River the pitch of the synclinal axis increases northeastward to 60°, accompanied by intense local folding and crumpling of the incompetent Grenville rocks in the vicinity of the synclinal trough. Minor fold axes are oriented both parallel to and plunging steeply toward the tectonic axis (Pl. 5). The latter attitude would suggest a certain amount of drag developed as a result of differential movements with strong horizontal components along the above-mentioned border faults. (*See* Faults; Fig. 10.)

It was not difficult to obtain dip and strike readings on the well-layered, though lithologically variable gneisses and schists of the complex. No discrepancy was observed in the orientation of foliation and presumably original bedding, and therefore no structural distinction was made between them. Foliation structures range from rather crude biotite parallelism in the highly granitized layers of the Canada Hill phase to excellent schistosity in the thin intercalated biotite and graphite schists of the Grenville series. Good gneissic banding permitted measurement of the hornblendic and augitic layers associated with the Pochuck diorite phase.

The determination of linear structure within the planes of foliation was hampered by the deeply weathered outcrops exhibiting the foliate facies of the Highlands complex. Linear alignment of such minerals as biotite, hornblende, pyroxenes, and graphite is probably more prevalent than would appear from the map (Pl. 5), since only reasonably certain linear orientations were recorded. Generally the linear elements seem to be elongated parallel to the minor as well as major fold axes thus conforming to the direction of intermediate elongation of folding deformation or the *b*-fabric (principal) axis of Sander (1930, p. 119). The study of lineation in the Highlands complex was limited to the orientation of linear components and minor fold axes insofar as such data were necessary to establish the regional structure and its relation to the Storm King intrusive. Thus a careful, detailed investigation would probably reveal lineation perpendicular to the fold axis—*i.e.*, parallel to the *a*-fabric axis of Sanders (1930, p. 119)—to be more common than in the few localities shown (Pl. 5, Fort Montgomery area and lower Doodletown Brook).

No attempt is made to give a genetic interpretation of lineation in these rocks, since too

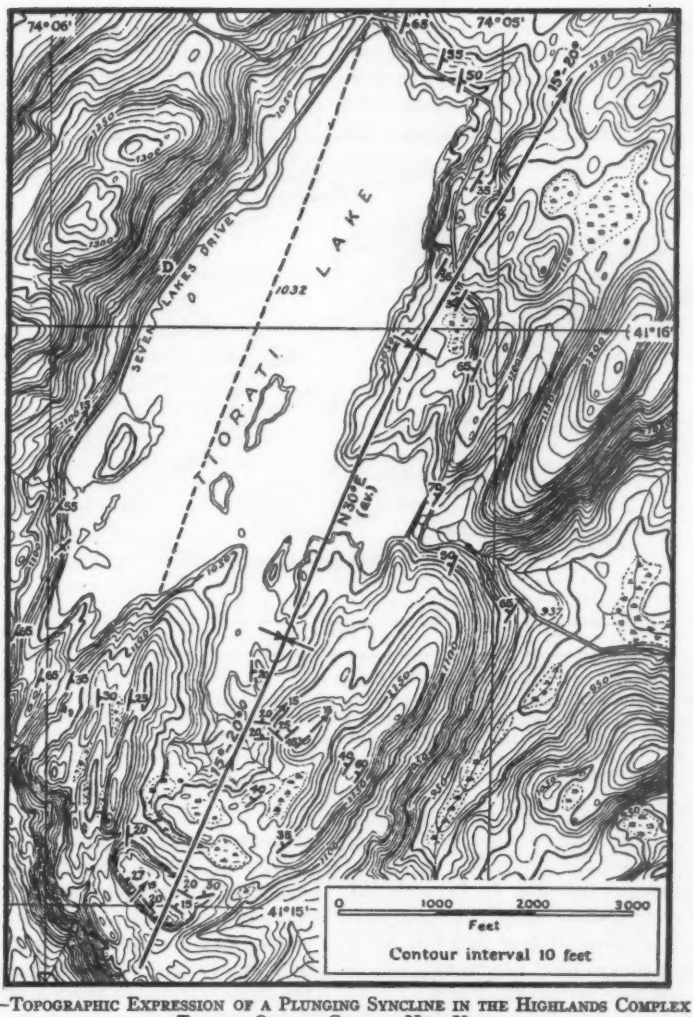


FIGURE 3.—TOPOGRAPHIC EXPRESSION OF A PLUNGING SYNCLINE IN THE HIGHLANDS COMPLEX AT LAKE TIORATI, ORANGE COUNTY, NEW YORK

Foliation indicated by strike and dip symbols. Linear flow structure of concordant Storm King granite bodies shown by arrows. Massive diorite (Pochuck phase) exposed at D. From Topographic Map of the Palisades Interstate Park, U.S. Geol. Survey, 1928, 1:12,000.

few observations were available. The evidence suggests that more than one process was responsible for lineations of similar position in space. The linear alignment of the mafic minerals in the granitic rocks of the Canada Hill phase probably was due to primary igneous flow controlled by the structure of the folded and metamorphosed Grenville strata and thus oriented parallel to the tectonic axes. Secondary deformation causing rotation of mineral grains due to intense folding (Cloos, E., 1946, p. 18) can account for lineation in the crumpled, wrinkled, and drag-folded Grenville gneisses and schists. Simultaneous stretching of the grains may have taken place due to tensional stresses in the direction of the tectonic axis

as indicated by the prominent cross fractures perpendicular to the fold axes of the Highlands complex. Furthermore, mineral elongation due to secondary flow deformation coupled with structure, aided in many places by the mobilizing effect of igneous intrusive phases (Fig. 5). Similarly *lit-par-lit* injections of the Pochuck and Canada Hill types were strongly deformed

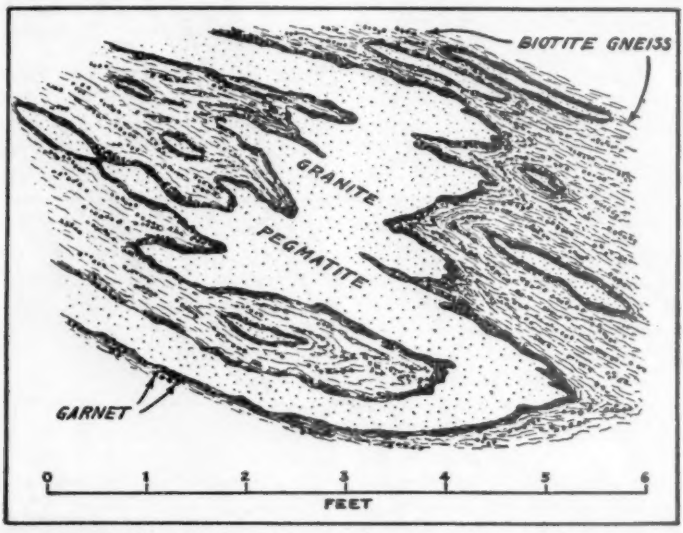


FIGURE 4.—PLASTIC DEFORMATION OF CANADA HILL AGE OR LATER IN HIGHLANDS COMPLEX

Concordant granite pegmatite (Canada Hill phase) intrusion into garnetiferous biotite gneiss (modified Grenville). Tectonic stress during or after emplacement caused intense flow deformation with boudinage structures. Garnet-biotite reaction rims suggest thermal metamorphism.

North shore of Roe Lake in Popolopen gorge, 75 feet west of dam, (08.82–00.63). Outcrop sloped 25° toward observer (south).

recrystallization was possible where the Grenville rocks appear to have been mobilized by the heat and volatile constituents of the invading igneous phases (Fig. 4). Certainly the origin of linear structure in the Highlands complex was much more involved than that in the Storm King granite where it is interpreted as a primary flow phenomenon exclusively.

Though the regional structure is clearly synclinal, close examination of the structure data (Pl. 5) shows considerable local variations in strike and dip indicating strong orogenic deformation of the metasediments. Particularly the thin-banded calcareous and micaceous facies are much crumpled and drag-folded, often producing isoclinal structures. Most of the minor axes, however, maintain the general trend and plunge of the larger structures. They are interpreted as adjustment by flow deformation of less competent beds to the regional stresses which produced the Highlands due to tectonic stresses during or after their emplacement (Fig. 4). Spencer *et al.* (1908, p. 2), for instance, explained at least part of the foliation in the Pochuck gneisses as the direct result of later injection of igneous material.

More than one deformational interval was evidently involved in producing the present structures. One of the most nearly constant features of Highlands geology is the remarkable control which the rock structures exerted upon even the earliest igneous phases. Colony (1921, p. 57–60) concluded that the magnetite ores were introduced into intensely metamorphosed and strongly folded Grenville strata. He also believed that "no folding of any magnitude has affected either the invaded complex or the ore bodies since the ores were deposited" (p. 61). The Bear Mountain area afforded no opportunity to study relations of magnetite deposits but did furnish many examples of evident deformation of thin *lit-par-lit* injected granite

pegmatites (Fig. 4). Even if these belong to the earliest igneous phase (Pochuck diorite of Berkey and Rice, 1919, or Pochuck gneiss of Colony, 1921), they were intruded either simultaneously with or subsequent to the emplacement of the magnetite. Consequently any deformation of these rocks suggests a later orogenic period which need not have been of the same magnitude or regional scale as the earlier tectonic stresses.

A relatively small portion of the Canada Hill phase appears as a massive biotite granite of the type described by Berkey and Rice (1919). A coarse pegmatitic facies with a mineralogical composition believed to be characteristic of this particular intrusive phase is far more prevalent and has been introduced *lit-par-lit*. Excellent examples of massive pegmatite sills, 10 feet and more in thickness, may be found on the south slope of the 600-foot hill south-southwest of Queensboro Furnace (06.42–99.55). The structural control exercised by the thin-banded, locally crumpled Grenville gneisses was so thorough that exposed contact surfaces of the pegmatite show a faithful image of the minor folds and crenulations of the country rock. Aside from some shear fracture zones traversing the intruded rocks, there is no evidence of folding or flow deformation subsequent to intrusion.

The limited influence of later orogenic forces is further shown by fractures and shear zones as well as the major fault structures trending mainly southwest to northeast which have been correlated with the Taconic and Appalachian revolutions (Colony, 1933, p. 27). Colony there reiterated his conviction that the later, massive igneous intrusions (particularly Canada Hill and Storm King granites) were little affected by either of these regional disturbances.

The major deformations responsible for the folded structure of the Highlands probably predated the earliest igneous rocks and were followed by other, progressively weaker tectonic disturbances. This concept accounts for the general lack of large-scale deformational effects, both megascopic and microscopic, in the granitic rocks of the Canada Hill phase and particularly of the still later Storm King granite. More detailed evidence concerning the latter is discussed elsewhere.

Bear Mountain is the intrusive core of a large down-folded area, the western and southern margins of which are roughly indicated by the arcuate trend of Turkey Hill, Summer Hill, and its southern extension (west side of Beechy

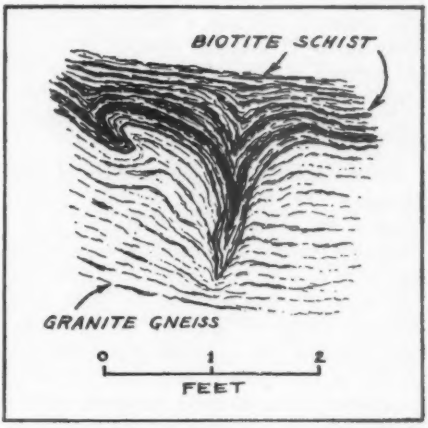


FIGURE 5.—MINOR FLOW FOLD IN GRENVILLE METASEDIMENTS MODIFIED BY CANADA HILL PHASE

Flow deformation of less competent biotite schist into biotite granite gneiss.

Perkins Memorial Drive (up), (07.42–98.88), 50 feet above road and 8 feet (true) below Storm King granite contact. Outcrop slopes 40° toward observer (southwest).

Bottom), West Mountain, The Timp, and Dunderberg (Pl. 1). Most of the eastern limb of this syncline appears to be absent probably because of displacement along the prominent Timp Pass-Hudson River fault. The Timp and Dunderberg, east of the main fault, quite likely represent part of this synclinal structure. Kemp (1888a) and others noticed the distortion and fracturing of the rocks along the Hudson River at the base of Dunderberg, and aerial photographs suggest that numerous shear and fault zones cut northeastward across the ridge—*i.e.*, roughly parallel to the main fault at Timp Pass. The abrupt change in ridge trend at the Pass from southeast in West Mountain to northeast in The Timp and Dunderberg is an added indication of the effect of faulting upon the folded structure. Unfortunately time did not permit adequate field investigation of this area.

The present gorge of the Hudson River in the

vicinity of Bear Mountain and Dunderberg marks the eastern limit of the syncline as exposed. The rather abrupt change of topography, lithology, and structural trend on the east shore of the Hudson River in Anthony's Nose and the mountainous territory to the east and southeast indicates a definite structural discontinuity (Thompson, 1936, p. 1838). On the strength of this evidence the writer has placed a tentative fault line following the bottom of the Hudson River in this locality (Pl. 1).

Closer examination of the structure of Bear Mountain (Pl. 5) reveals increasing disturbance of the normal synclinal attitude of the Highlands complex southeastward approaching the Timp Pass-Hudson River fault. Since the core of the syncline was filled by Storm King granite concordantly intruded into the Highlands complex, an overlying and underlying portion of the country rocks may be distinguished and will be referred to as top and bottom of the syncline, respectively. The top of the structure north of US 6 and the Bear Mountain bridge approach is a rather simple shallow syncline. At the northern end of Hessian Lake, however, the normal syncline seems to pass into an equally gentle anticlinal warp continuing to the shores of the Hudson River.

A structural problem is presented by the narrow "tongue" of Highlands complex rocks projecting westward into the body of the Storm King granite intrusive south of Hessian Lake. It finds topographic expression in a narrow, curving ridge rising from 200 feet elevation at the foot of the ski jump west of Bear Mountain Inn to 650 feet some 600 yards to the west-southwest. Absence of outcrops across the large playing field and parking area south of the Inn makes adequate interpretation of the structure difficult. Recently, construction work along the west side of the playing field temporarily exposed some bedrock which unquestionably belongs to the Highlands complex and indicates a distinct broadening of the base of the "tongue" southeastward without any noticeable change in structure. The writer assumes that the "tongue" is a part of the overlying country rock. Yet this concept introduces complications in trying to explain the gently folded Highlands gneisses in the "tongue", striking approximately E.-W. and dipping 30° to 40° N. This structure is incongruous with respect to the attitude of the Highlands complex some 1000 feet to the east. Three possible interpretations come to mind:

(1) Faulting might account for this change in structure, though no evidence of faulting was discovered. The lack of any topographic expression of such a fault is significant, because the fault would have to traverse the 160-foot rock bench known to be covered by only a veneer of glacial drift and man-made fill. Subaerial erosion which has been active perhaps since early Mesozoic times would surely have left a tell-tale mark on the present topography.

(2) A more plausible explanation involves sagging of a sliver of country rock into the Storm King magma after it was partially stoped from the top of the syncline. The large amount of heated igneous material compared to the relatively small volume of the enveloped rock sliver could well have mobilized the latter sufficiently so that the "tongue" could not maintain its original position. The fulcrum of rotation would then have to be sought in the vicinity of the easternmost outcrop of the "tongue", 300 feet southwest of Bear Mountain Inn. Again, structural corroboration is lacking. Longitudinal and cross joint systems characteristically found in the Highlands complex are excellently developed as smooth, plane, and continuous surfaces with regular spacing. Shear fractures and slickensides are absent. Structural attitude of the rather massive, garnetiferous, and highly silicified gneisses is quite constant for the entire length of the exposure (about 150 feet nearly perpendicular to the strike). No crumpling or drag folding was observed. "*Augen*" of granitic material within thin intercalated biotite schist layers are flattened in the plane of foliation and elongated parallel to the regional tectonic axis (N. 45° E., 40°). The enveloping schistose material exhibits prominent lineation of the same spatial attitude produced by parallel arrangement of fiber like mineral grains, mainly biotite and sillimanite ("lineare Faserung", H. Cloos, 1921). These deformational features appear to be definitely linked to the tectonic history of the Highlands complex prior to the intrusion of the Storm King granite.

(3) Local overturning of the gneisses in the "tongue" as a result of intense Grenville folding (earlier than the Pochuck igneous phase) must be considered. A number of rather poor and discontinuous outcrops of the Highlands complex along the eastern margin of the parking area south of Bear Mountain Inn suggest considerable local disturbance of the gently warped top of the regional synclinal structure. A small, but well-preserved overturned fold at the northwest corner of the Bear Mountain Post Office and garage (09.53–99.01), for instance, shows that the co-ordinates of the minor fold axis closely coincide with those of the regional tectonic axis. The prominent longitudinal joint truncating the exposure also maintains the orientation of the regional system. This minor feature then could well be repeated on a larger scale, and the "tongue" would represent the overturned portion. The "tongue" may also be more extensive and continue some distance into the mass of the Storm King intrusive beyond its apparent termination against the base of the steep eastern flank of Bear Mountain. However, any connection between the overturned structure and the presumably normal attitude of the rocks in the bottom of the syncline is pure speculation.

The writer favors the last explanation considering recorded evidences of similar intense folding of the Grenville gneisses in the Highlands both on a large and small scale. Cook (1868, p. 58) shows an overturned, asymmetrical and nearly isoclinal syncline in the cross-section of the Hurdstown mine in northern New Jersey. Colony (1921, p. 32) describes the structure of the great ore body beneath Mine Hill at Franklin Furnace, New Jersey, as "bent back upon itself to form a long trough with sides of unequal height". He furthermore considers it more likely that this folding preceded the emplacement of the ore body—*i.e.*, probably in Grenville time. He also (p. 100) refers to the structure in the Forest of Dean Mine, 3 miles northwest of Bear Mountain, as follows:

"The ore apparently occupies the trough of a synclinal fold, overturned to the northwest, and pitching toward the northeast."

At the bottom of the Bear Mountain syncline, the most striking feature is the pronounced downward curvature of the structure along the lower Perkins Memorial Drive (up) which is quite out of harmony with the gently concave corresponding portion of the top. A similar discrepancy, though on a smaller scale and partly inferred, is indicated along the west flank of The Torne. This phenomenon is believed to have been caused by the sagging of the structure concurrent with the intrusion of the Storm King granite and is treated more fully in the chapter dealing with the origin of the Storm King granite.

The previously discussed disturbance of the synclinal structure toward the Timp Pass-Hudson River fault zone is well shown by a sharp anticlinal fold south of the "tongue" and by irregular dips and strikes and much crumpling and shear fracturing of Grenville rocks along the lower Doodletown Brook valley.

Structures in the Storm King Granite

General statement.—The Storm King granite was described by Berkey and Rice (1919, p. 56) as possessing "... a marked, but crude gneissoid structure". Although no further reference was made to the mineral structure, the use of the term "gneissoid" suggested the presence of mineral parallelism as a result of primary igneous flow. The same authors also stated (p. 37) in connection with a discussion of the Highlands granites:

"Even the simpler granites very rarely have perfectly massive structure. They are almost universally streaked or crudely banded in a way which suggests at once some form of flowage... by far the more common structure is strictly conformable with the regional structure or trend".

The results of the detailed structural study of the Storm King granite at Bear Mountain agree with the very general concept expressed by the earlier investigators.

Strong weathering effects on most of the prominently exposed rock ledges generally limited reliable measurements of internal mineral structures to reasonably fresh road cuts (Pl. 5). Excellent recent exposures at significant

and critical locations were furnished by the construction of the Perkins Memorial Drive which was opened to traffic in 1934.

The Storm King granite in this region exhibits rather prominent lineation on nearly all exposures. Hornblende crystals and crystal aggregates of rather stout prismatic habit may be arranged in both linear and plane-parallel structures. This "matchbox" shape of the hornblende components generally results in linear traces showing insufficient variation in elongation, width, and continuity to permit a clear recognition of linear and/or planar alignment of the crystal elements. The presence and attitude of linear structure becomes evident only when a "structureless" or even-granular outcrop surface, oriented essentially perpendicular to the axes of the linear components, can be found.

The construction and study of transparent models presenting all available field data for a particular outcrop (Lowe, 1946a) proved of considerable help in determining the spatial arrangement of the linear components in doubtful cases. Fourteen such models representing critical outcrops of questionable structure were made and studied in detail. Since the accuracy of the results obtained by this method compared favorably with direct field measurements, it was not thought necessary to make any distinction between these two types of data in plotting them on Plate 5.

A graphic-statistical method (Lowe, 1946b) was employed to increase the accuracy of determining co-ordinates of linear structure from field data. Again the stubby shape of the linear hornblende components was the principal reason for the frequently crude results of direct field measurements.

Contacts.—Bear Mountain offers an unusually good opportunity to study the relations of the Storm King granite intrusive into the country rocks of the Highlands complex owing to contact exposures aggregating some 5000 feet. This feature is unique in the Hudson Highlands where extensive weathering and erosion have reduced the less resistant Highlands complex rocks to topographic lows. Contacts, therefore, are generally at or near the base of the granite ridges where they are commonly covered by glacial drift and alluvium.

Four thousand feet of the lower granite contact is almost continuously exposed along the southern flank of the mountain. The exposed contact crosses the lower Perkins Memorial Drive (up) at 910 feet elevation (07.30-98.95) (Pl. 3). Trending southeastward, it parallels the down-drive until it is concealed by an intersecting curve in the road at 830 feet elevation (08.08–98.53). About 250 feet east and 30 feet below road level it emerges again marked by a prominent overhang which can be traced through the woods for another 1500 feet. At 670 feet elevation (08.62–98.51) the bedrock disappears beneath heavy glacial drift. Other contact exposures, though rarely more than 100 feet in length, were found in several critical localities along both the upper and lower margins of the granite intrusion (Pl. 5). Perhaps the most significant of these established the exact southern border of the "tongue" of Highlands complex rocks projecting into the Storm King granite mass southwest of Bear Mountain Inn (09.00–98.98).

Although exposures comprised only about 10 per cent of the total length of contacts mapped, the margins of the Storm King granite intrusive could be determined with considerable accuracy. Exposures were commonly adequate within a short distance either side of the inferred contacts, or erosional features indicating such lines of weakness were present. Particularly where Storm King granite had invaded the Grenville metasediments of the Highlands complex, the difference in resistance to erosion of the two rock types was responsible for significant geomorphic forms (Fig. 6).

A distinct break in slope with local development of granite cliffs 20 to 30 feet high marks the lower contact of the intrusive between Seven Lakes Drive (09.15–98.27) and US 9W (09.68–98.60). The narrow stream valley east of US 9W paralleling the highway was characteristically developed along the upper contact with the Highlands complex. Where the Storm King granite intruded the weaker, thin-bedded, schistose and in some places calcareous Grenville rocks, the northeast-dipping structure encouraged subsurface drainage rather than surface run-off and produced the linear swampy depression between US 6 (08.40–00.50) and the north end of Hessian Lake (08.85–00.05).

This crescent-shaped lake hugging the base

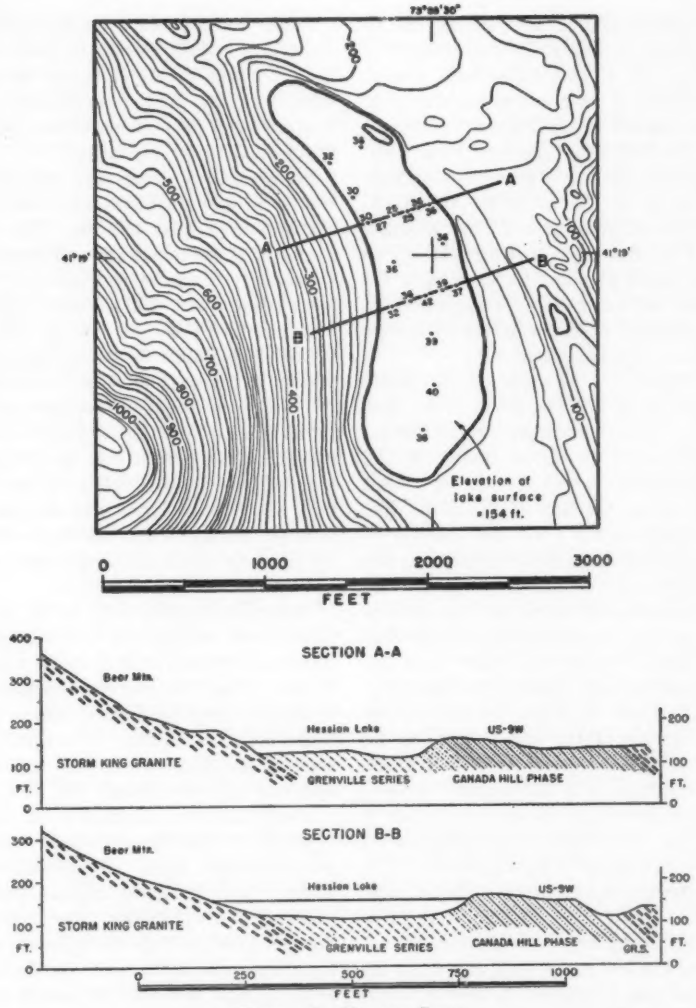


FIGURE 6.—HESSIAN LAKE

Topographic map and geologic cross sections. Depth soundings by courtesy of A. K. Morgan, General Manager, Palisades Interstate Park, Bear Mountain, New York.

of the steep Storm King granite flank of Bear Mountain is interpreted as an extension of this contact-line depression, somewhat deepened by glacial excavation of the weathered Grenville rocks. Glacial striae along the path following the west shore of the lake (09.05–99.65) trend S. 30° E.—*i.e.*, parallel to the northern portion of the crescent. The margins of the Lake basin appear to have been determined by rocks more resistant to weathering and erosion than the Grenville beds. The west and south shores are composed of Storm King granite, while the east shore is rimmed by the Canada Hill granite phase of the Highlands complex. Only near the northern tip of the lake do Grenville rocks extend continuously northwestward to US 6 (Pl. 5).

The strike of foliation in the country rocks

conforms remarkably to the curving eastern shore and seems to support the interpretation of Hessian Lake as a contact-line basin. Soundings establishing a maximum depth of only 42 feet were obtained by the Engineering Department of the Palisades Interstate Park Commission at Bear Mountain and were obligingly made available to the writer by Mr. A. K. Morgan, General Manager and Chief Engineer (Fig. 6). Two cross sections of the lake following the principal sounding traverses show the shallow lake basin and make the assumption of glacial excavation of the weathered Grenville contact beds quite plausible.

A good exposure of the upper granite contact on the east shore, 500 feet north of the Inn, falls in line with the contact trend following the opposite shore to the northwest (Pl. 5). This might justify drawing the inferred contact diagonally across the lake at this point. Yet, such inference would pose the problem of accounting for the origin of the southern portion of the crescent-shaped lake basin which would have to transgress such a contact. The uniform depth of the lake bottom across this hypothetical line throws considerable doubt upon this interpretation, since the greater erosional resistance of the Storm King granite would result in a definite rise of the lake bottom on the granite side of the contact. As discussed previously, existence of a north-south trending fault zone is not tenable in view of the lack of field evidence indicating such a structure across the rock bench south of Hessian Lake.

On the strength of geomorphic evidence, then, the writer believes the contact of the Storm King granite closely follows the southern margin of the lake. The few granite exposures along the eastern shore are interpreted as part of a narrow, curved, sheetlike portion of the main intrusive body. This concept also fits into the structural scheme of the Highlands complex in this locality.

In contrast to the rather satisfactory delineation of the intrusive margins at Bear Mountain, the contacts along the flanks of The Torne are, for the most part, buried under glacial drift. Except for one poor outcrop on the eastern slope north of Ft. Montgomery road (08.10–01.06) the location of inferred contacts may well be in error by some hundreds of feet.

Linear flow structure.—The most evident and constant structural feature of the Storm King granite is the linear alignment of the hornblende crystals. This phenomenon is largely responsible for the pronounced streaky lineation apparent in most exposures.

Plate 5 shows clearly the remarkable uniformity of spatial orientation of this structure. Figure 9 emphasizes this fact. Fifty measurements of linear structures scattered over approximately 1 square mile of exposed Storm King granite were plotted on a Schmidt net and contoured (Billings, 1942, p. 119).

Objections might be raised concerning the validity of results based on so few observations, but each set of data plotted represents the mean for an outcrop. Thirty-five of these values were determined from road cuts by graphic-statistical solution (Lowe, 1946b) on the basis of field data obtained from 6 to 10 exposure surfaces per outcrop. These road cuts were at least 50 feet long and 15 feet high, and some were several times as large.

The resultant value of N. 40° E., 40° for the mean linear structure in the Bear Mountain intrusive, then, is believed to be valid within narrow limits. This attitude coincides with the trend of the regional tectonic axis of the Highlands complex syncline as determined from regional structure data (Pl. 5) and measurements of numerous minor fold axes.

Hornblende, the principal dark constituent of the Storm King granite, appears to have crystallized early, when the magma was still mobile. Well-developed crystal outlines are therefore not uncommon. The later feldspar crystals show no such development or dimensional alignment in the field or under the microscope. Quartz, generally the last to crystallize, is likewise anhedral. Although the larger quartz grains are elongated roughly parallel to the linear structure of the hornblende, there is no comparable space-lattice orientation. This anomaly is considered unrelated to the origin of the hornblende lineation, and a tentative explanation is given in connection with petrofabric analysis.

Since the mineral structure of the Storm King granite is predominantly linear, involves mainly hornblende crystals of early magmatic origin, and has uniform dimensional and crys-

tallographic orientation parallel to the regional tectonic axis, it is interpreted as the result of primary igneous flow under control of pre-existing deformation structures—*i.e.*, upward at an angle of about 40° from the horizontal in the general direction S. 40° W., or nearly parallel to the axis of the Bear Mountain syncline. The flow was probably laminar and relatively quiet. Locally irregular and wavy flow lines suggest some turbulent flow.

The writer inclines toward the belief that linear flow structure oriented parallel to the tectonic trend of the country rocks is characteristic of the Storm King granite or its equivalent throughout the Highlands. He has found support for this in exposures of this rock in the Hudson Highlands of Rockland and Orange counties (Fig. 1). The literature, too, contains significant supporting records. Thompson (1936, p. 1836) speaks of flow lines in the Storm King granite at the northwestern gateway of the Hudson River gorge through Storm King and Breakneck Mountains and "elsewhere in the Highlands". Since Balk (1936, Pl. 20) did not differentiate the various granites and gneisses of the Highlands in Dutchess and Putnam counties, it is not known which of the numerous, generally northeastward-plunging flow lines represent linear orientation of hornblende in Storm King granite or its equivalent.

Considering the inadequate lineation data reported from the Hudson Highlands, it is interesting to note that geologists studying the New Jersey Highlands, early recognized this feature. Wolff (1893, p. 362–364) described linear structure in a massive hornblende granite gneiss and called it "pitch structure". He considered it a structural characteristic of this rock type, noted its parallelism to the regional northeasterly plunge, and explained its origin as due to primary crystallization.

The writer feels confident that the lithological, mineralogical, and structural evidence contained in Wolff's report indicates that the granite gneiss described is the equivalent of the Storm King granite of the Hudson Highlands. This contention is bolstered by reference to the linear structure of the dark components (principally hornblende) of the Byram granite gneiss in the Highlands of northwestern New Jersey and eastern Pennsylvania (Bayley, 1941). The correlation of this rock type with the Storm King granite was established by Berkey and Rice (1919, p. 140). Bayley observed (p. 150) that the long axes of the pencil-shaped dark minerals were aligned parallel to the axes of near-by magnetite ore bodies and that these in turn were conformable with the regional plunge of the rock structure (p. 56). He also showed (p. 50) that a lighter variety of the Byram granite gneiss was characterized by a "lack of dark pencils" and only a "slight linear structure". This agrees with the contact phase of the Storm King granite at Bear Mountain and certainly implies that the linear structure in the Byram granite gneiss, as in the Storm King granite, is produced mainly by the dark minerals.

Bayley (1910, p. 126) regarded the linear structure of the gneisses in the New Jersey Highlands as the result of flowage of viscous magma and of crystallization of some of the minerals under strains produced by the flowage. Similar references to the origin of linear structure in the Byram granite gneiss may be found in Spencer *et al.* (1908, p. 2) and Bayley *et al.* (1914, p. 9).

Platy flow structure.—The linear hornblende elements in some places tend to align in parallel planes of limited extent. Thus the linear traces on an exposure surface oriented perpendicular to the dip of such planar structure appear as parallel discontinuous streaks. This surface pattern is practically indistinguishable from the one produced by most outcrop truncations of Storm King granite possessing linear flow structure only. Yet sometimes the increase of biotite content in the near-contact facies of the granite accentuates this plane-parallel arrangement of the mafic minerals and gives the rock a more nearly foliated appearance.

Planar structures, with few exceptions, can be observed only in the marginal zones of the granite and in the relatively narrow sill-like portion exposed east of highway US 9W near Bear Mountain Inn, where they are invariably oriented parallel to the nearest contact surface of the intrusive. Flow lines are generally present as well, maintaining their characteristic orientation parallel to the regional tectonic axis. In many instances this linear parallelism could not be ascertained due to intensive weathering of the outcrops or the lack of sufficient dark

mineral grains. On the other hand, the remarkable sensitivity of erosional processes to oriented, though megascopically obscure, rock fabrics is clearly shown in some places. Northeast of the Bear Mountain swimming pool (approximately 09.50–99.70), for instance, a series of parallel hogback ridges of Storm King granite (contact phase) 10 to 15 feet high trends N. 50° W., parallel to the strike of the structure. The gentler eastern slopes (40°) are developed parallel to the platy flow structure of the granite, while the western slopes follow a system of excellent cross joints dipping 55°. Foot paths crossing these ridges utilize small cross valleys and sags evidently controlled by the prominent longitudinal joint system, N. 20° E., 70° W.

The occurrence and orientation of these hornblende-biotite layers are expectable along the confining walls of a concordant igneous mass intruded under viscous laminar flow conditions (Fairbairn, 1937, p. 111). Consequently they are interpreted as platy flow structure or flow layers. In fact, the attitude of flow layers was an indicator of the approximate location and trend of concealed contacts.

Support for primary flow origin of the mineral alignment in the Storm King granite can be established by application of the principal structural criteria for distinguishing primary from secondary gneisses as listed by Balk (1937, p. 151).

The limited geological literature on the Hudson Highlands does not seem to relate the occasionally "banded structure" of the gneissoid Storm King granite to its clearly stated magmatic mode of emplacement (Berkey and Rice, 1919, p. 34, 70).

Reports from the New Jersey Highlands, on the other hand, yield more specific information. Wolff (1893) recognized not only the well-developed linear structure in the massive hornblende gneisses (presumably equivalent to the Storm King granite) near Hibernia, N. J., but also mentioned that it may suddenly become "a plane-parallel or laminated one" (p. 364) or may often be "combined with a plane of foliation" (p. 363). Despite the evidence, admittedly suggestive or primary flow, the author favored the concept of metamorphism and recrystallization contemporaneous with folding and *without* fusion to explain the observed mineral alignment. Later Wolff (1895, p. 19) regarded similar granitic rocks occupying extensive areas in the Franklin Furnance, N. J., region as undeniably intrusive. An accompanying report on the geology of Jenny Jump Mtn., N. J., by Westgate (1895, p. 33), however, still considered their "banded" structure as probably of secondary origin by calling at least a part of the granite gneisses "metamorphosed granite". Spencer *et al.* (1908, p. 2) finally linked these same structures to the primary origin of the rocks. Similar opinions, some specifically referring to the Byram granite gneiss, were expressed by Bayley (1910, p. 126; 1942, p. 47, 50) and Bayley *et al.* (1914, p. 5).

Joints

Compilation and study of joints.—Since study of joints[5] in an area such as the complex crystalline terrane of the Hudson Highlands is dependent on adequate measurement and recording of data, the writer adopted the following arbitrary limitations in the belief that they would not jeopardize justifiable interpretation of the results (Pl. 4):

(1) Strikes and dips refer to *joint sets* rather than individual joints.

(2) They constitute the *mean attitude* of each set per outcrop, determined from a representative number of individual joint measurements.

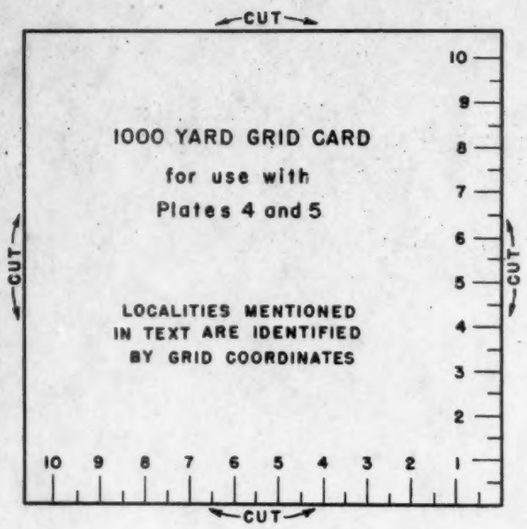
(3) Single joints and those lacking definite indications of systematic grouping were not recorded. Likewise no measurements were obtained from poor outcrops.

(4) Only the *principal joint systems* were recorded and their variables evaluated on the basis of an arbitrary field classification:

I Most prominent and best developed; smooth plane surfaces; regular spacing at relatively close intervals (6 inches to 8 feet); large and continuous; numerous joints per set.

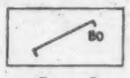
II Prominent and well developed; smooth to rough, generally plane

[5] The term "joints" includes both open fractures ("Spalten") and closed fractures ("Klüfte") as distinguished by Cloos, H. (1936, p. 214).

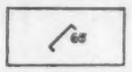


LEGEND

Principal joint sets, with strike and dip
(Bedding joints, sheeting, and occasional joints lacking sytematic attitude are not shown on this map)



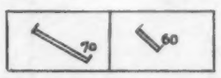
Group I.
(Most prominent and best developed; smooth, plane surfaces; regular spacing, 6 in. to 8 ft.; large and continuous; numerous joints per set)



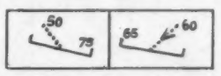
Group II.
(Prominent and well developed; smooth to rough surfaces; irregular spacing, 6 in. to 8 ft.; sometimes discontinuous; fair number of joints per set)

Group III.
(Least prominent, but well developed; rough, curving surfaces; irregular, wide spacing, more than 8 ft.; large and continuous; few joints per set)

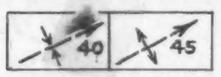
Vertical joint sets



Dense jointing, spacing less than 6 inches; often slickensided

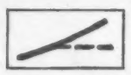


Slickensided joints, with bearing and pitch of striae; arrow indicates relative displacement of hanging wall where known



Syncline Anticline

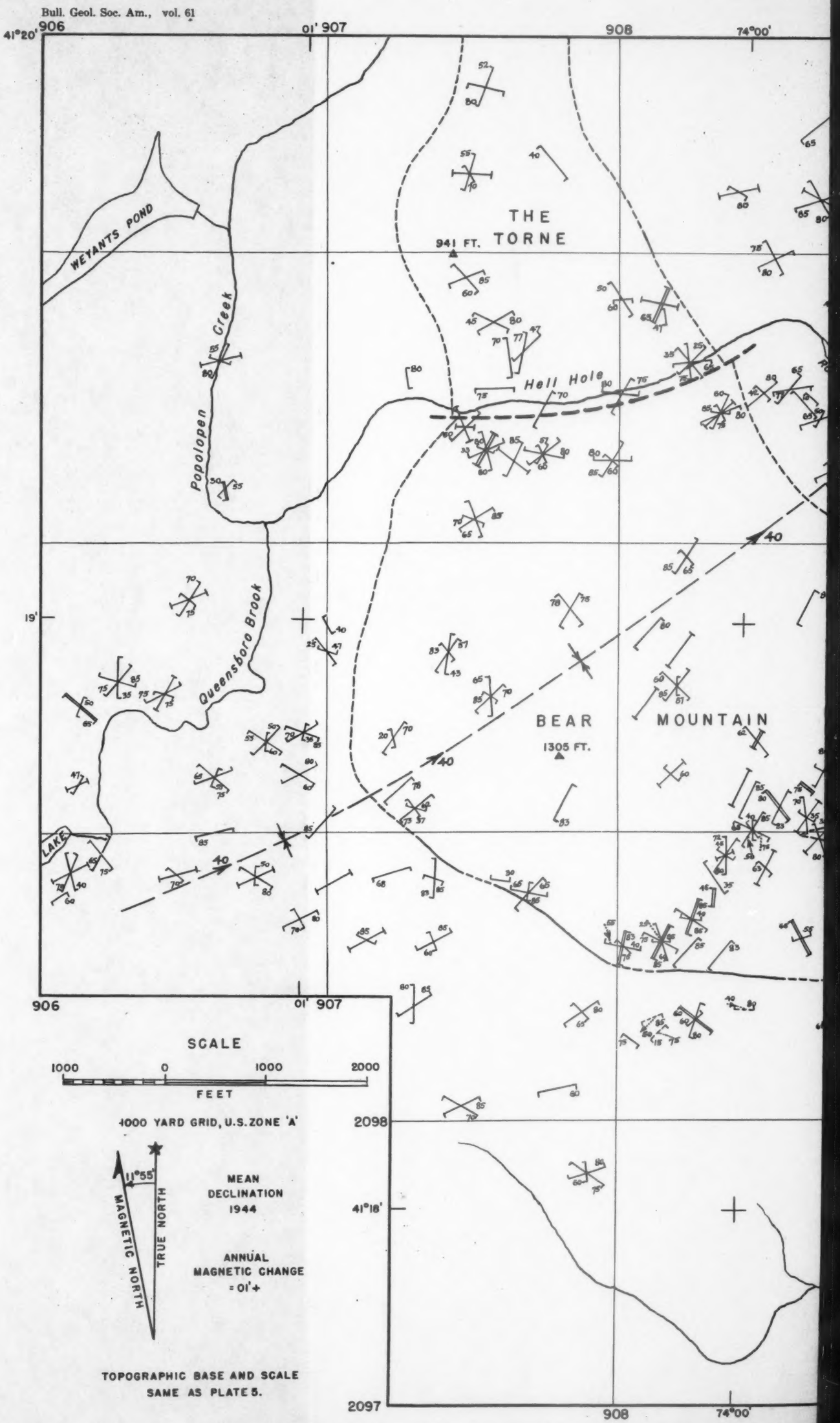
Tectonic axes, with bearing and plunge



Faults, observed and inferred

Contacts of Storm King granite, observed and inferred

Granite lies *within* area bounded by contact lines; compare geologic map, Plate 5



JOINTS AT BEAR MOUNTAIN, NI

LEGEND

…t sets, with strike and dip
…ting, and occasional joints lacking
…de are not shown on this map)

Group I.
…d best developed; smooth, plane
… spacing, 6 in. to 8 ft.; large and
…rous joints per set)

Group II.
…ll developed; smooth to rough
…ar spacing, 6 in. to 8 ft.; sometimes
…ir number of joints per set)

Group III.
…but well developed; rough, curving
…ar, wide spacing, more than 8 ft.;
…ous; few joints per set)

Vertical joint sets

…g, spacing less than 6 inches;
…often slickensided

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…hanging wall where known

Syncline Anticline

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…ts of Storm King granite,
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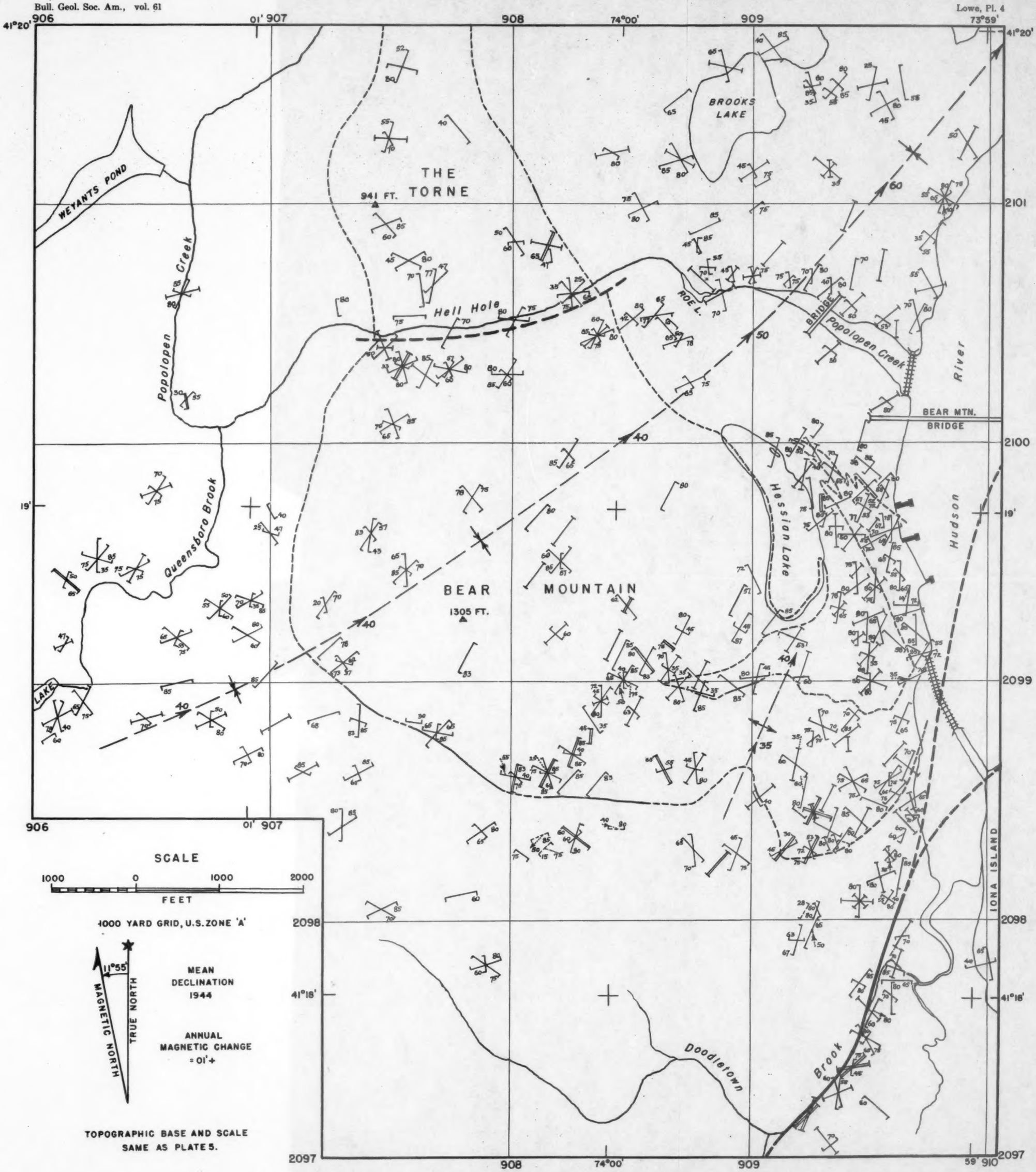
…hin area bounded by contact lines;
…re geologic map, Plate 5



SCALE

1000 0 1000 2000

FEET

4000 YARD GRID, U.S. ZONE 'A'

11°55'

MAGNETIC NORTH

TRUE NORTH

MEAN DECLINATION 1944

ANNUAL MAGNETIC CHANGE = 01'+

TOPOGRAPHIC BASE AND SCALE SAME AS PLATE 5.

JOINTS AT BEAR MOUNTAIN, NEW YORK

surfaces; irregular spacing; limited in size and sometimes discontinuous; often local in character; fair number of joints per set.

III Least prominent, though well developed; mostly rough and curving (wavy) surfaces; irregular and wide spacing (more than 8 feet); large and continuous; only few joints per set; this type is generaly not shown on Plate 4.

(5) Sheeting (Billings, 1942, p. 128) produced by exfoliation was not recorded, since this type of jointing was found to be a near-surface phenomenon and invariably oriented parallel to the existing topography.

(6) "Bedding" joints parallel to the foliation were omitted, although they are not uncommon in the gneisses and schists of the Highlands complex. Their development is essentially local and increases with fissility of the rocks involved. Inclusion of their ever-changing spatial orientations in the statistical study would tend to obscure the preferred orientation of the significant principal joint systems in relation to the well-defined axis of regional deformation.

(7) Local zones of densely crowded parallel joints (fractions of an inch to 6 inches apart) are indicated on Plate 4. Slickensides were noted in the field, but their presence was recorded only where the attitude of striae could be determined.

To facilitate recognition and interpretation of the principal joint systems on a regional scale, the map plot (Pl. 4) is supplemented by contour diagrams on pole projections on an equal-area net (Billings, 1942, p. 118). Joints in the Highlands complex and in the Storm King granite are treated separately (Fig. 7). The dominant regional structure features—*i.e.*, trend of tectonic axis and attitude of linear flow structure respectively—are plotted for reference. In addition the mean spatial orientation of joint sets corresponding to pole maxima of the contour diagrams is shown on standard stereographic projections (Fig. 8).

Types and distribution of joints.—The principal joint systems in the country rocks of the Highlands complex are closely similar in orientation to the Storm King granite intrusive (Fig. 8). Equally comparable are their relative frequencies as expressed by the per cent concentration maxima of Figure 7. Yet this phenomenon is not too surprising in view of the parallel orientation of the regional tectonic axis and the linear flow structure in the concordantly intruded Storm King granite. Thus whatever the nature and cause of the force which affected the two types of terrane, the directions along which they acted and the resulting fracture patterns have similar spatial orientations.

The three main joint systems at Bear Mountain can be classified on a purely geometrical basis with respect to the uniform trend and plunge on the regional tectonic axis (N. 45° E., 40°). In order of their relative frequency, they are longitudinal, diagonal, and cross joints. Their dip angles are generally steep (70°–90°); all numerical co-ordinates given are mean values derived from statistical study (Fig. 7).

The longitudinal system strikes N. 30°–36° E. and has nearly vertical dips. These joints are by far the most prominent both in development and frequency (more than 9 per cent).

Diagonal joints, striking approximately N. 5° W. and dipping 73° W., are much less conspicuous in the field but rank second in frequency distribution (4–6 per cent).

Cross joints, N. 50°–55° W., 60°–68° S., are the rarest (2–4 per cent in the Highlands complex and 4–6 per cent in the Storm King granite) but not the least prominent in field development. In many localities they are rated as belonging to Group (I) of the writer's field classification (Pl. 4).

The contour diagrams (Fig. 7) were obtained by plotting all measurements of joint sets regardless of their relative prominence in the field, as expressed by the writer's joint groups (I), (II) and (III). Therefore the concentration maxima are significant only insofar as they indicate the mean attitude of preferred joint orientation and the relative per cent distribution of such orientation. Comparative field evaluation as recorded on Plate 4 is an essential corollary to statistical results when attempting a genetic interpretation of joints; also additional field data dealing with joint fillings, rela-

tive age, evidence of subsequent movements, etc. have to be taken into consideration.

Joints in the Highlands complex.—Statistical investigation (Figs. 7, 8) supplemented by careful study of field occurrence as plotted on Plate 4 revealed some rather remarkable features pertaining to the development and distribution of joints in the Highlands complex.

(1) Great predominance of longitudinal joints (*S* joints of Cloos), parallel to the tectonic axis, both in excellence of development and frequency of occurrence.
(2) Relatively minor frequency of well-developed cross joints (*Q* joints of Cloos).
(3) Occurrence of conjugate sets composed of longitudinal and cross joints.
(4) Absence of *systematic* conjugate shear joints which can be related to the known direction of maximum regional compression.
(5) Preferential development of a *single* set of diagonal joints roughly bisecting the northwest and southeast quadrants between the longitudinal and cross joints.

Regional orogenic deformation was undoubtedly the major cause of jointing in the intensely metamorphosed, folded, and faulted Highlands complex terrane. The writer agrees with Colony (1921, p. 57–60) that the present folded structure is largely the result of Precambrian, presumably Grenville, orogeny. The resulting deformation structures were so well defined that they consistently exercised structural control over successive igneous invasions. Under these conditions it is difficult to see how the Appalachian orogeny could have been solely responsible for the present Highlands complex joints as claimed by Appleby (1942) for the correlated crystalline rocks of the northern New Jersey Highlands. Appleby's contention that jointing of earlier origin was "obliterated" by the Appalachian deformation must be rejected on several grounds.

The Storm King granite at Bear Mountain shows no evidence of secondary foliation or any internal fabric pattern indicating marked regional tectonic influence. Although it is possible, at least in the Hudson Highlands, that this granite could have been intruded in early Paleozoic times, it would be difficult to imagine its emplacement after or during the Appalachian orogeny. Consequently if the later deformation failed to affect this youngest member of the crystalline "basement", it certainly could not have destroyed any existing joints in the Highlands complex. Furthermore regional faulting in the Bear Mountain area suggests that the tectonic trends of the Precambrian and Paleozoic orogenies can be differentiated.

In some instances exposures permit determination of the relative ages of joint systems. Along the eastern shore of Hessian Lake (09.19–99.84), for instance, two equally prominent longitudinal joint sets have the co-ordinates N. 10° E., 85° W. and N. 35° E., 85° W. The first set is truncated by the second and is therefore later. In other words, the older northeast-trending joints might be interpreted as indicative of Precambrian deformation, while the more nearly northerly strike of the later set could be correlated with Paleozoic orogeny. The evidence is circumstantial, to be sure, but fits too well into the picture of regional faulting and topographic trends to be entirely coincidental.

Their relatively small angle of divergence (25°), however, certainly would not support the concept of "obliteration" of earlier fracture systems even under the influence of a hypothetically all-powerful Appalachian orogeny. On the contrary, such later deformation forces acting roughly in the same direction as the original stresses should have emphasized the earlier joint pattern and produced subsequent movement along such directions of existing rock weakness. Much of the slickensiding of fracture surfaces at Bear Mountain is believed to have been produced in just this manner. Appleby (1942, p. 29) also suggested that subsequent (Appalachian) joints may have followed earlier joint patterns so that no difference is recognizable between the earlier and later fractures. This may well be true in portions of the western New Jersey Highlands where the structural trends of the Precambrian crystallines and the infolded and infaulted Paleozoic rocks appear to be parallel.

The possible existence of primary joints in the Highlands complex must also be considered. During the Pochuck diorite and Canada Hill granite phases great quantities of igneous material were introduced largely conformable with

the regional structures. Stresses set up by volume changes due to thermal effects and (re) crystallization certainly produced fracture systems. Such primary joints, however, are admittedly difficult, if not impossible, to distinguish from the similarly oriented earlier deformation fractures. Except for late-stage granite pegmatites, the various igneous phases seemed to have been guided by the folded structure rather than the associated fracture systems. This fact coupled with the observation that, "The joint planes everywhere cut the injections and the parent rock without change of direction", led Appleby (1942, p. 28) to conclude that "jointing, therefore, was not associated directly with the magmatic invasions of the area.... The stresses developing out of the magmas did not give rise to joints sufficiently strong to survive subsequent diastrophic action".

Field evidence at Bear Mountain contradicts this. Figure 4 of Plate 3 illustrates a Highlands complex inclusion in the marginal portion of the Storm King granite. The set of well-developed joints, confined to the inclusion, is part of the regional longitudinal system in the nearby parent rocks and therefore pre-Storm King in age. The larger joint on the left appears to be related to the Storm King intrusion, since it arises in the granite and does not transgress the contact zone 8 feet below (Plate 3, fig. 2, left). This fracture cuts across the inclusion and has the same orientation as the earlier joints in the country rocks of the Highlands complex. Its primary origin is also indicated by the fact that many similar longitudinal joints elsewhere in the granite mass are filled by magmatic end-stage products such as aplite, epidote, and chlorite.

It is not unreasonable to conclude, therefore, that jointing in the Highlands complex is predominantly of tectonic origin and probably reflects Precambrian rather than Paleozoic deformation.

(1) The remarkable development of near-vertical longitudinal joints oriented perpendicular to the direction of maximum regional compression (NW.-SE.) appears to be unusual.

The preferred statistical orientation of these joints compared to any other system of joints at Bear Mountain might be a condition peculiar to this local downfolded structure. In the absence of data from the Hudson Highlands as a whole, Thompson (1936, p. 1841) has shown that the frequency of longitudinal joints (striking NE.-SW.) along the Hudson Gorge is subordinate to that of cross joints (striking NW.-SE.). Thompson[6] made no distinction between joints in the Highlands complex and in the Storm King granite, but his field data clearly indicate that the frequency distribution determined by him is valid for both formations in the limited region under investigation. A further basis for comparison with the writer's statistical conclusions is given by Thompson's joint map (not published) which included only the principal joint systems and indicated their relative frequency by relative lengths of lines.

The detailed joint plot at Bear Mountain (Pl. 4) also suggests a certain differential development of longitudinal joints in relation to the local structure. These joints seem to be best developed near the major synclinal axis particularly in the Fort Montgomery area (northeast of Popolopen Creek) and east of Queensboro Lake. There they commonly occur as the only principal joints.

Their dominant though less obvious development in the Highlands complex around Bear Mountain Inn might be associated with the rather intense local deformation where the axes of minor isoclinal and overturned folds are parallel to and part of the regional tectonic trend.

The occasional slickensiding of longitudinal and other joints at Bear Mountain indicates movement, possibly in Taconic or Appalachian time, but sheds no light on the cause of the original planes of rock weakness.

Tensional stresses in response to a release of pressure (Billings, 1942, p. 125) evidently could not have played an important role in this region where active compression is presumed to have existed more or less continuously from Grenville to Appalachian times. Yet rupture under tension appears to be indicated when we consider that the longitudinal joints at Bear Mountain are generally smooth, plane, exten-

[6] The writer is indebted to Professor Thompson for the opportunity to examine unpublished data and maps upon which his statistical diagram (Fig. 5, p. 1841) was based.

sive surfaces quite regularly spaced and as a rule not slickensided. Since these joints tend to be relatively more numerous and better developed near fold axes, tensile stresses caused by "stretching over the crests" (Nevin, 1942, p. 140) during folding deformation could have had some influence on their development. Also, in the discussion on foliation in the complex, the sharp downward flexure of the Bear Mountain syncline below the granite intrusion has been compared with the relatively gentle curvature of the structure in the rocks overlying the granite. The writer suggested an actively sinking region as a corollary to the Storm King intrusion to account for this disharmonious structure. Sagging of the structure into the emptying reservoir of the Storm King magma would have affected the deeper portions of the synclinal trough relatively more than its limbs. Thus renewed tensional stresses across the axis of folding would account for additional longitudinal joints.

Shearing of the type represented by fracture cleavage (Billings, 1942, p. 230), however, cannot be ruled out. Figure 5 of Plate 3 shows a thin bed of biotite-hornblende gneiss between rather massive, granitic layers in the Highlands complex. A set of high-angle dip joints appears to have developed only in the relatively incompetent gneiss. Since the exposure is near the "nose" of the Bear Mountain syncline, this joint set parallels its axial plane and can therefore be classified as longitudinal. While fracturing in this manner is quite feasible locally where basic schistose and gneissic Grenville rocks predominate, it does not explain the origin of the principal longitudinal joints in the Highlands complex which cut across competent and incompetent beds alike (Pl. 3, fig. 3).

Linear structures in the foliate facies of the Highlands complex have been produced by elongated crystals of biotite, hornblende, graphite, sillimanite, etc. Their long dimensional axes generally parallel the axes of minor and major folds. This lineation might have been the result of rotation of grains around the fold axis perhaps aided by stretching or flowage in the direction of intermediate elongation of the fold (parallel to the fold axis) as shown by E. Cloos (1946, p. 37). Closely spaced linear elements of this type could have caused planes of weakness in the rock fabric parallel to the axial plane, thus cutting across the foliation. But in that case one would expect such joint surfaces to be somewhat uneven and to exhibit prominent minute linear crenulations parallel to the fold axis. This condition was rarely observed in the Highlands complex.

Several possible mechanisms have been suggested to account for the striking development of longitudinal joints in the Highlands complex at Bear Mountain. Yet no one line of reasoning appears to offer a wholly satisfactory explanation particularly in view of inadequate factual data on similar joint patterns in other regions of comparable lithology and structure.

The writer found few pertinent joint observations in the literature, with one notable exception. Osborne and Lowther (1936) found prominent longitudinal joints oriented parallel to the regional tectonic axis in Grenville metamorphics but made no attempt to explain their origin in terms of deformation mechanics.

H. Cloos (1936, p. 229) and Nevin (1942, p. 140) refer to longitudinal master joints in gently folded, massive sedimentary rocks. The several papers explained such joints either as the result of tensional forces set up by arching of the folds or as due to shearing under compression. The second explanation seems unconvincing in view of experimental evidence of rupture of rocks under shearing stress (Billings, 1942, Chap. 6) pointing to the formation of conjugate shear joints arranged symmetrically about the direction of maximum compression.

(2) The problem of accounting for the prominent *cross joint system* is much less controversial. There is abundant evidence (H. Cloos, 1936, p. 227; Billings, 1942, p. 125) that tensile stresses parallel to the fold axis (direction of intermediate elongation) develop cross joints. In addition, flexing of the tectonic axis at Bear Mountain undoubtedly accentuated this stress condition. The axial plunge of the Bear Mountain syncline is 40° NE., exceeding the average value for the Highlands by some 10°. Furthermore the angle of plunge steepens northeastward to 55°–60° in the Fort Montgomery area indicating a downward bending of the axis.

This structural feature is also suggested by the scattering of cross-joint concentrations (3-maximum) on the contour diagram (Fig. 7,

H-1). Since tensional joints develop perpendicular to the fold axis, the range of joint dips from 40° to 90° apparently reflects local changes in the axial plunge.

Although Cloos believes that both systems were the result of shearing, he admits that many occurrences of this type cannot be explained in this manner. This writer sees no

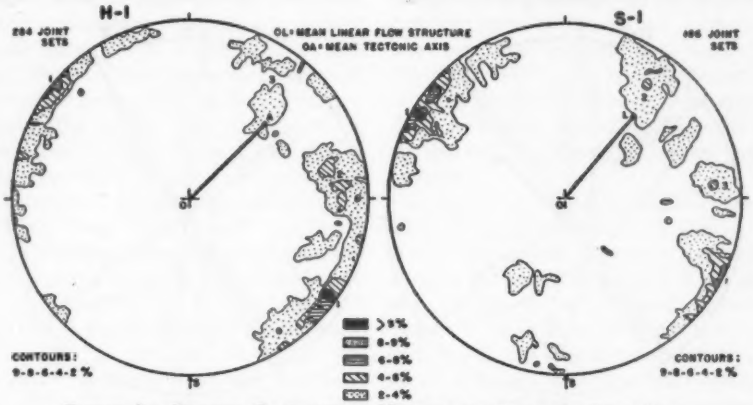


FIGURE 7.—CONTOUR DIAGRAMS OF NORMALS TO PRINCIPAL JOINT SETS

H-1 Highlands complex, S-1 Storm King granite. Maxima are numbered in order of decreasing concentrations. Lower hemisphere projection on Schmidt equal-area net.

Comparatively little significance is attached to the low statistical frequency of cross joints at Bear Mountain (Fig. 8, H-2). The purely relative value of 2–4 per cent expresses the extraordinary concentration of longitudinal joint orientation. Thompson's frequency diagram (1936, p. 1841) probably gives a more nearly correct picture of the strong development of cross joints in the Hudson Highlands.

(3) Many exposures of the Highlands complex exhibit only two principal joint systems frequently representing longitudinal and cross joints. In such cases we may speak of conjugate joints particularly when there is no evidence that they were formed at different times or in response to more than one set of stresses. The term conjugate is used in the sense of H. Cloos (1936, p. 229) who implies that "Kluftpaare" are formed essentially simultaneously during a single interval of regional deformation. The angle between such joint pairs may be 90° or less, with 60° as the approximate minimum. Cloos also distinguishes between equally and unequally developed pairs. He makes special mention of "unequally developed, perpendicular, conjugate joints" which are frequently composed of longitudinal and cross joints. This is the exact condition found at Bear Mountain. obstacle to considering tensile stresses as the cause of the cross joint system. Consequently, if we can apply the term conjugate according to Cloos, the longitudinal system was probably due to tension as well.

Billings (1942, p. 127) suggested that vertica joints perpendicular to each other may be produced as conjugate shear planes by a couple, assuming horizontal movements parallel to the strike of vertical border faults. Such movements along parallel faults may well have taken place in the structural development of the region. There is reason for doubting, however, that conjugate shear planes intersecting at right angles ever form in nature (Billings, 1942, p. 108).

The limited scope of this investigation failed to turn up certain evidence considered essential for adequate interpretation of joint origin. A good deal more needs to be known, for instance, concerning the relative age of the various joint systems. Only the most detailed and painstaking study could hope to obtain such data.

(4) Statistical studies and results (Fig. 7, H-1) give no indication of the existence of other sets of conjugate joints. Field data (Pl. 4) suggest the presence of possible conjugate sets

intersecting at angles of less than 90° and commonly interpreted as shear joints.

However, there appears to be no systematic arrangement. Not only do the angles of intersection vary, but joint orientations with respect to principal stress directions are likewise inconstant. The direction of maximum compression (SE.-NW.) bisects the acute as often as the obtuse angle between these joint pairs. This fact alone throws some doubt on the validity of referring to such paired systems as conjugate shear joints.

In any event, a curious anomaly exists in this strongly folded region of known principal stress directions; typical conjugate shear joints supposedly associated with folding deformation seem to be absent.

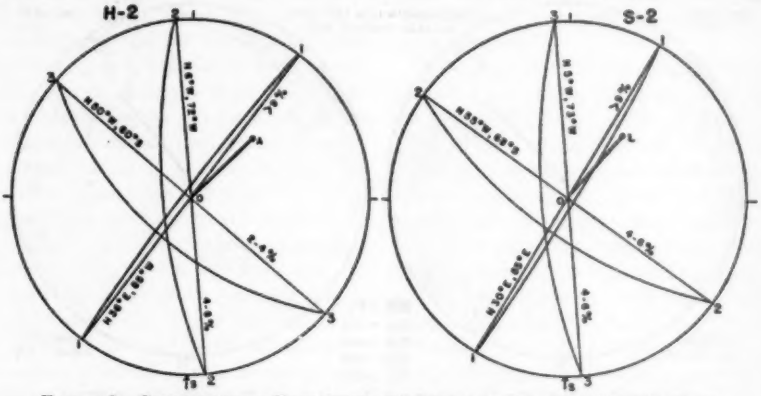


FIGURE 8.—STEREOGRAPHIC PROJECTION OF MAXIMA OF PRINCIPAL JOINT SYSTEMS

H-2 Highlands complex, S-2 Storm King granite. Systems plotted represent mean values of maxima from contour diagrams (Fig. 7). Numbers and per cent concentrations correspond. Lower hemisphere projection on Wulff net.

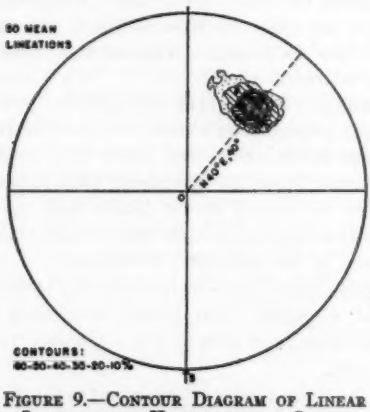


FIGURE 9.—CONTOUR DIAGRAM OF LINEAR STRUCTURE OF HORNBLENDE IN STORM KING GRANITE

Each lineation measurement represents mean bearing and plunge for an entire outcrop. Lower hemisphere projection on Schmidt equal area net.

(5) Another distinctive feature is the prevalence of single sets of diagonal joints bisecting the angle between the principal longitudinal and cross joints. Their preferred orientation (Fig. 8, H-2) appears to be N. 6° W., 72° W., with a frequency of 4–6 per cent (Fig. 7, H-1) or appreciably greater than that of cross joints. There is little reason for calling them shear joints in the absence of a requisite set of complementary joints arranged symmetrically about the direction of maximum compression. It is suggested, therefore, that this diagonal system may be tear joints analagous to tensional feather joints (Billings, 1942, p. 124) produced by a couple.

The borders of the Bear Mountain syncline have been affected by movements along two major faults trending somewhat obliquely to the tectonic axis and gradually converging northeastward (Pl. 1). The Timp Pass-Hudson River border fault on the southeast is the more prominent and probably originated in Grenville time. It is interpreted as a high-angle thrust fault from the southeast. The Crown

Ridge fault appears to have disturbed only the conformable relations between the Highlands complex and the Storm King intrusion along the northwestern limb of the syncline.

North of Fort Montgomery the ridges developed on Highlands complex rocks have a northerly trend (Pl. 1) which changes quite abruptly along the fault to the northeast trend of the Crown Ridge and Bare Rock composed of Storm King granite (Fig. 10). This fault, however, seems to die out in the granite west of Brooks Lake where the intrusive is conformable with the trough of the Bear Mountain syncline. The Crown Ridge fault then is post-Storm King and probably was activated in response to thrust movements along the Timp Pass-Hudson River fault zone. In that case the horizontal components of such movements could produce stress couples as indicated in Figure 10. Furthermore the locally prominent transverse folding at Fort Montgomery (*see* discussion of foliation and folds) suggests that the stipulated fault movements had appreciable horizontal components.

In accordance with this interpretation the diagonal joints in the Highlands complex would be late tension joints presumably due to Paleozoic rather than Precambrian deformation.

In view of the inconclusive interpretation of the clear-cut joint pattern at Bear Mountain, the work of Osborne and Lowther (1936) at Shawanigan Falls, Quebec, is noteworthy. The authors made a detailed petrofabric study of joint orientation in Grenville metamorphics and in a granulite phacolith representing successive deformation intervals with strongly divergent tectonic trends (N. 45° W. and N. 22° E. respectively). In each case, well-developed sets of vertical strike (longitudinal) and dip (cross) joints were oriented parallel and perpendicular to the respective tectonic axes. Joint orientations appeared to be continuous and constant regardless of the strike of foliation in any one locality. Small maxima of joint concentrations suggested the presence of E.-W. joints. The authors did not consider this evidence strong enough to stipulate another distinct system. Yet the east-west strike with respect to the regional tectonic trends is quite analogous to the orientation of the diagonal system at Bear Mountain.

The resulting regional joint pattern is comparable with that at Bear Mountain. The authors did not explain the origin of the longitudinal system, although the cross joints were interpreted as tension joints. The indication of east-west joints on the contour diagrams may have been more apparent than real owing to the rather small number of joints plotted (60 to 70 per diagram), but Osborne did suggest that they may represent E.-W. tension joints which are generally widespread throughout the Laurentians.

Joints in the Storm King granite.—The great similarity of statistical joint distribution in the Storm King granite to that in the Highlands complex has been mentioned. A minor discrepancy appears to be the slightly greater frequency of cross joints (4–6 per cent) in the granite, but this may be more statistical than actual, since only 186 joint systems were plotted to obtain the Storm King granite contour diagram (Fig. 7, S-1), while the results of the Highlands complex (Fig. 7, H-1) were based on 284 measurements.

The significant features of joint orientation and frequency in the Highlands complex are equally applicable to the Storm King granite, adding supporting evidence that the granite is concordant. The principal stress in an intrusive is maximum elongation in the direction of movement. At Bear Mountain this stress was applied parallel to the northeast-plunging synclinal axis, parallel to the direction of intermediate elongation as well as perpendicular to that of maximum shortening of the earlier folding deformation. Presumably the Storm King granite was intruded under compressive stress due to continued, though perhaps attenuated, orogeny (Fig. 14). Thus the principal strain axes of the intrusion had essentially the same spatial orientation as two of the major directions of strain arising from the orogenic deformation of the country rocks. Therefore one would expect the resulting fracture systems in the Highlands complex and the Storm King granite to be comparable, if not identical.

It has been suggested that regional Paleozoic deformation was responsible for some of the principal joint systems in the Highlands complex. The same should hold for some of Storm King granite joints, but petrologic evidence

points to quiet intrusion and fractures are the only secondary structures in the granite. These might indicate either limited intensity of post Storm King orogeny or shallow depth of the crystalline rocks at the time of such later deformation thus producing mechanical rupture rather than flowage and recrystallization.

On the whole, the joints in the granite are believed to be primary fractures directly associated with emplacement. Longitudinal, cross, and diagonal joints contain in places typical magmatic end-stage products including pegmatite, aplite, epidote, chlorite, sericite, and quartz. Most of these joint fillings are confined to fractures within the granite mass and undoubtedly had their source in the Storm King magma. Although only one pegmatite dike was observed crossing the contact, several were found in the country rocks some distance from the nearest granite exposure. In each case the characteristic mineral assemblage (hornblende-microcline-microperthite-quartz) left no doubt as to their origin. Furthermore there is no other possible source, since the only post-Storm King magmatic activity in the Highlands involved relatively minor intrusions of rapidly chilled basalt (camptonite) dikes.

A certain number of the dominant longitudinal joints show features suggested by Balk (1937, p. 34–36) as characteristic of primary fractures of this type. Many of them are filled with epidote, veneered with sericite, and impregnated with limonite. Occasionally they also served as channels for aplite and pegmatite dikes. Their tendency to "develop zones in which they are closely spaced" (Balk, 1937, p. 35) is especially noticeable at Bear Mountain (Pl. 4). Epidotization and impregnation with iron oxides is not uncommon in these zones, but exessive disintegration and kaolinization is not general. Slickensiding of the closely spaced joint planes is common and indicates that later stresses were relieved by movements along directions of greatest primary weakness.

To account for the primary origin of longitudinal joints in the Storm King granite appears no less difficult than to explain their formation as due to folding deformation in the Highlands comple.. However, the linear structure of the hornblende in the granite is the principal element of weakness in the rock fabric whose other constituents show little systematic spatial orientation. Also flow layers are near-contact phenomena and not characteristic of the intrusive as a whole. The writer, therefore, quite agrees with Balk (1937, p. 35) that "... fractures that embrace the flow lines form relatively easily".

It is impossible to judge the relative proportion of primary longitudinal fractures to those formed subsequently. Theoretical reasoning might favor primary origin, but field observations show that primary mineralization at Bear Mountain is much less common in longitudinal than in cross and diagonal joints.

Concerning the origin of cross joints in the Storm King granite the writer agrees with H. Cloos (1921) and Balk (1937, p. 27–33). Their experimental and field data establish rupture under tension created by elongation of the intrusive body in the flow direction as the principal mechanism.

By far the most common orientations of filled fractures at Bear Mountain belong to the cross joint system which likewise shows the greatest variety of hydrothermal mineral products. All the end-stage minerals and dike rocks listed above occupy cross joints or coat the joint surfaces.

Later basic dikes appear to have had the same preference for occupying cross joints. Thompson (1936, p. 1837) noted that "the granite adjacent to the dikes shows closer spacing of the joints than elsewhere". He related this feature to the greater weakness of the rock in zones of closely spaced joints. The writer's field observations corroborate this point, but the question may be raised why much more frequent zones of dense longitudinal joints at Bear Mountain are not associated with a proportionately larger number of basaltic dikes. Balk (1937, p. 36) suggests that similar conditions in various granite intrusions may be explained by lateral (tectonic) pressures keeping longitudinal fractures closed while causing distention of cross joints.

H. Cloos (1925, p. 33) that cross joints in igneous bodies are of primary age only, while longitudinal and diagonal joints may be primary or secondary. The writer has shown that the decreasing dips of cross joints in the Highlands complex at Fort Montgomery may be related

to the downward bending of the synclinal axis. The Storm King granite here is well below the surface, and the flexure, therefore, cannot be dated in relation to the emplacement of the granite. If it occurred in post-Storm King time, however, certainly some cross joints in the granite could have formed in response to this secondary deformation.

Diagonal joints (Balk, 1937, p. 37) are represented by the statistical concentration of steep fractures striking at angles of approximately 45° to the trend of the flow lines (Fig. 7, S-1; fig. 8, S-2). Probably most are primary, because many are occupied by aplite dikes and quartz stringers, and the walls of many are coated with chlorite. Only one set of diagonal joints (mean orientation: N. 5° W., 73° W.) appears to have developed preferentially, just as observed in the Highlands complex.

Balk (1937, p. 37) explains primary diagonal joints as shear planes, referred to a predominantly horizontal (compressive) stress. Although this concept requires conjugate systems, he mentions that there may be only one diagonal set, the second having been suppressed, but he offers no reasons for this partial development (p. 38).

He suggests, however, that diagonal joints formed during the stage of incipient crystallization of the top of an intrusion serve a similar purpose as cross joints by tending to elongate the rock mass in the direction of the trend of flow lines (p. 39).

This was observed by the writer in only one instance. Two hundred feet east of the Seven Lakes Drive (09.38–98.26) a prominent diagonal joint (N. 20° W., 80° W.) in the Storm King granite near its lower contact cuts across a series of quartz stringers oriented parallel to the platy flow structure (N. 75° E., 30° N.). The quartz bands are displaced on the hanging wall of the joint showing offsets of 1 inch or less on the near-vertical exposure surface trending roughly east-west. Flow lines indicate magmatic movement at an angle of about 25° upward from the northeast. The net effect of this displacement is to distend the rock mass slightly in the direction of linear flow as illustrated by Balk (1937, p. 37, Fig. 14, right).

Tensional forces produced by a couple could also have caused diagonal fractures. When the cooling magma became viscous enough to permit rupture in the crystallizing mass, a condition analogous to the formation of crevasses in glaciers could have existed (Billings, 1942, p. 124). Magmatic flow in the thick trough portion of the intrusion must have been relatively more rapid than in the restricted, curving channels of the eastern limb where frictional resistance was greater. Rather uniform trend and plunge of flow lines (N. 40° E., 40°) indicate parallel linear motion throughout the intrusive, and tear fractures of this type would have orientations similar to the observed diagonal joint system.

This mechanism also requires the development of opposing tear joints toward the western limb of the synclinal intrusion. The granite body north of Hell Hole, however, thins more gradually so that the frictional differential would be less. Consequently one would not only expect fewer fractures but would also anticipate their strike to be more nearly parallel to the trend of the flow lines.

Joint measurements along Hell Hole and on The Torne (Pl. 4) support this hypothesis. The usual set of diagonal joints appears to be replaced by a system of steep joints trending north of east. This condition is reflected in the contour diagram (Fig. 7, S-1) by a minor concentration of joint poles corresponding to mean joint co-ordinates of N. 60° E., 80° S.

The minor twin maxima in the southwest quadrant of the contour diagram (Fig. 7, S-1) represent joint measurements of E.-W. to N. 55° W., 35°–50° N. These are fractures parallel to the exposed lower granite contact due to cooling of the marginal portion of the intrusive.

Faults

Regional development.—The early reports of the New Jersey Geological Survey contain many references to the fault structures in the crystalline Highlands of New Jersey. Cook (1883) and Britton (1886) distinguished extensive longitudinal faults trending northeast and dipping northwest with overthrusts to the southeast, and shorter cross faults. The latter appear to be the more numerous in this region and were interpreted as gravity faults with horizontal offsets to the southeast.

Berkey and Rice (1919) found abundant evidence of longitudinal thrust faulting in the Hudson Highlands, but reported overthrusting to the northwest. Some of these strike faults in Peekskill Hollow were interpreted as due to block faulting probably in Triassic time or earlier (p. 79). The authors also indicated that Precambrian faulting produced brecciation and shear zones in the crystallines which are now healed and can be easily overlooked in the field owing to their lack of topographic expression. Paleozoic (Taconic and Appalachian) and later faulting, on the other hand, is generally reflected in the present topography and can be detected more readily (p. 76). Little mention was made of cross faults which seem to play such an important role in the New Jersey Highlands. North-south trending block faults were shown slicing across the narrow band of Paleozoic rocks in Peekskill Hollow (map) and penetrating the crystallines for some distance on either side, but these are evidently not comparable in direction or development to the typical cross faults of the New Jersey Highlands.

Timp Pass-Hudson River fault.—The writer has applied this geographical description to one of the most prominent longitudinal faults in the Hudson Highlands (Pl. 1). Berkey and Rice (1919) mapped this fault along the Hudson Gorge from Garrison to Anthonys Nose where it crosses the river west of Iona Island and follows the lower Doodletown and Timp Brook valleys to the western margin of the West Point quadrangle (74° 00′) meridian. The writer established its continuation along the upper Timp Brook and across Timp Pass. There is also evidence that John C. Haff and John W. Butler, Jr.[7] traced this fault for about 2 miles southwestward from Timp Pass nearly to the southern margin of the Highlands.

The trend of the Hudson River Gorge between West Point and Bear Mountain Bridge (Anthony's Nose) is undoubtedly closely related to this fault (Thompson, 1936, p. 1836). The lower Doodletown Brook is also subsequent upon this structure and flows in a narrow gorge exhibiting abundant evidence of fault deformation in its steep rock walls. The Timp Brook evidently developed as a fault-line tributary of the Doodletown Brook. But, except in the upper reaches, direct field observations of faulting in Timp Brook are lacking owing to the heavy overburden of glacial drift and detrital material derived from the abrupt flanks of West Mountain and Dunderberg to the south. Timp Pass, on the other hand, provides an excellent example of the effectiveness of differential erosion in the Highlands. Removal of intensely sheared and crushed rocks in the fault zone has produced a steep-walled notch 1200 feet wide at the top and 250 feet deep in the otherwise highly resistant granite ridge. This geomorphic feature, accentuated by the near-vertical east wall of the Pass, is a distinctive landmark visible for many miles.

The writer believes that the Timp Pass-Hudson River fault, at least south of Bear Mountain, originated in Precambrian time as a high-angle overthrust from the southeast. The folded structures of the Highlands complex certainly predated the emplacement of the Storm King granite and probably also the intrusive phases of the Canada Hill granite and Pochuck diorite. The evident disturbance of the broad synclinal structure south and southeast of Bear Mountain Inn must then be the result of strong local Precambrian deformation. While the intensity of folding, crumpling, and shearing of the Highlands complex increases southeastward toward the fault, similar deformational structures are almost absent east of this zone. Also minor, mostly isoclinal folds in this area are consistently overturned to the northwest. Therefore overthrusting from the southeast in Precambrian time was the responsible tectonic agent.

The surface of dislocation is thought to dip steeply to the southeast, as suggested by the relative narrowness of the zone of most intense deformation exposed in Doodletown Brook. Furthermore densely crowded, slickensided shear fractures and crush zones trend essentially parallel to the rock walls of the gorge and have vertical to steep easterly dips. Their attitude indicates the general orientation of the thrust plane since they represent surfaces of some of the fault movements. Striae on the slickensided fracture walls are of little help in determining

[7] Field map of the southeastern corner of the Schunemunk quadrangle (1934) found among the effects of Professor R. J. Colony.

the direction of thrusting. They are generally smooth grooves and plunge in all directions. Where relative movements can be determined hanging walls seem to be displaced upward as often as downward. At the mouth of Doodletown Brook where the fault crosses US 9W no systematic arrangement of joint is apparent. Instead curving, slickensided fractures divide the rocks into irregular blocks which evidently suffered much rotational displacement. Movements in this fault zone, then, were most complex and probably persisted through more than one period of geologic time.

Berkey and Rice (1919, p. 76 and map) refer to a Precambrian fault crossing the Hudson River northeastward from the northern tip of Iona Island. Significantly enough the authors linked this structure with the Timp Pass-Hudson River fault at the mouth of Doodletown Brook. This establishes a roughly linear fault zone from Timp Pass northeastward—*i.e.*, parallel to the trend of Precambrian deformation in the Hudson Highlands.

The writer has intimated that the present fault structures could not have been produced by movements in Precambrian time alone. At Timp Pass the fault has truncated the concordantly intruded Storm King granite of West Mountain. Vertical fault striae on the west wall of the Pass clearly reveal that the relative displacement of this side (foot wall) was downward. Thus the Highlands complex of the Timp evidently represents a portion of the country rocks underlying the sheetlike granite intrusion which was upthrust into its present discordant position. It is not unreasonable to assume that the Storm King granite in West Mountain and Dunderberg was emplaced as a single sheet conformable with the synclinal structure of the region, very much in the manner of the Bear Mountain intrusion. Strong fault movements, therefore, must have occurred in post-Storm King (possibly Paleozoic) time to "break the back" of the syncline in Timp Pass.

A healed zone of brecciation in the Storm King granite is exposed on the Seven Lakes Drive between (09.32–98.35) and (09.35–98.57). Linear parallelism of the elongated granite fragments and principal fracture fillings indicate disruptive movements in the direction N. 25° E. This trend conforms to the strike of the Timp Pass-Hudson River fault 1500 feet to the southeast. Despite the circumstantial nature of the evidence, it is thought that brecciation was caused by post-Storm King movements along this fault. No other major fracture zone exists in the vicinity which could account for the intense disruption of the granite.

In the Hudson River gorge north of Doodletown Brook the fault trends generally N. 22° E., representing a distinct departure from the presumably Precambrian trend of the Hudson Highlands as indicated by the synclinal axis at Bear Mountain (N. 45° E.) and the Iona Island fault of Berkey and Rice (N. 45° E.). Is it coincidence, then, that the ridge crests north of Fort Montgomery (Pl. 1; Fig. 10) assume a trend parallel to the Hudson River fault? The writer is likewise impressed by the recurrence of the same north-northeast trend of faults and Grenville beds east of the Hudson River (Berkey and Rice, 1919, map). In Dutchess and Putnam counties (Balk, 1936, P. 20) the same structural orientation can be observed in the Highlands crystallines as well as in the Cambro-Ordovician Hudson River pelite.

This evidence suggests that the divergent trend of certain structural units in the Hudson Highlands may be related to the effects of Taconic or Appalachian deformation. Thus the evident lateral warping of the synclinal axis south of Highland Falls (Fig. 10) is believed to have been produced in Paleozoic time by thrusting along the Timp Pass-Hudson River fault zone.

E. Cloos (1940) demonstrated the axial divergence of "Appalachian" tectonic trends and Pre-"Appalachian" (Precambrian) trends of the basement in southeastern Pennsylvania and Maryland. He included in the basement rocks the Pickering, Pochuck, and Byram gneisses which are characteristic constituents of the Pennsylvania and New Jersey Highlands. Speaking of the "Appalachian" trends he noted: "This trend is almost north-south in New York" and "... large portions of the Precambrian basement seem to be passively involved" (p. 855).

Despite the absence of Paleozoic rocks in the Hudson Highlands investigated by the writer, the idea of "passive involvement" of the Precambrian crystallines in Paleozoic orogenies, to

the extent of faulting and consequent change of tectonic trends of the affected rocks, seems quite plausible.

Crown Ridge fault.—The writer did not make a detailed field study of the Crown Ridge fault (Pl. 1; Fig. 10), but certain deductions can be made from areal geology and geomorphic features. This fault truncates the folded structure of the Highlands complex south of Highland Falls but appears to strike parallel to the structural trend of the Storm King granite.

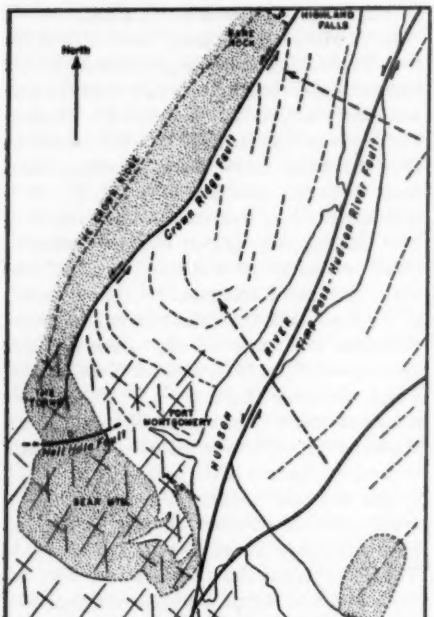

North
Babe Rock
HIGHLAND FALLS
Crown Ridge Fault
THE TORNE
FORT MONTGOMERY
Hell Hole Fault
BEAR MTN.
HUDSON RIVER
High Peak - Hudson River Fault
0
1
2
3
MILES
THRUST FAULTS, WITH HORIZONTAL COMPONENTS OF MOVEMENT
PROBABLE DIRECTION OF THRUSTING
NORMAL OR TEAR FAULTS WITH LARGE VERTICAL COMPONENT, D=DOWNTHROWN SIDE
TRENDS OF PRINCIPAL JOINT SYSTEMS (STATISTICAL MEAN VALUES FROM FIG. 9). LENGTH OF LINES APPROXIMATELY INDICATES RELATIVE FREQUENCY OF JOINTS PER SYSTEM
TREND OF RIDGE CRESTS
STORM KING GRANITE, WITH AND WITHOUT CONTACTS

FIGURE 10.—FAULT AND JOINT PATTERNS IN BEAR MOUNTAIN REGION

Horizontal components of thrust faults are hypothetical. Trends of ridges closely follow strike of folded structure.

Overthrusting from the southeast along the Hudson River fault must have crowded the strongly folded, well-foliated and therefore relatively mobile Highlands complex west of the Hudson River against the massive granite in the Crown Ridge. Since the Storm King granite here is evidently part of the concordant synclinal intrusion at Bear Mountain and The Torne (Pl. 1), the southeast-dipping surface of the thick sill provided an ideal gliding plane. That the granite effectively resisted these thrust movements is indicated by the fact that the Crown Ridge fault seems to die out west of Brooks Lake against the eastern granite flank of The Torne. This is inferred from geomorphic evidence alone, because of the heavy cover of glacial drift in this locality. There is no indication that this fault maintains its southwesterly trend across The Torne. Likewise the possibility of its curving southward along the east flank of The Torne must be eliminated, because there is no evidence of it in the bedrock exposed by Popolopen Creek in Hell Hole gorge (Pl. 1; Fig. 10).

Consequently the freedom of movement enjoyed by the Highlands complex block between The Crown Ridge and the Hudson River increased to the northeast. This condition should have produced a decided upward rotational displacement in the vicinity of Highland Falls which would account for the westward deflection of the syncline in this area. Also, increasing rotational movement should be accompanied by decreasing angles of plunge of the affected rock structures. The foliation plunges 25° to 30° at Highland Falls (Thompson, 1936, Fig. 9, p. 1845). This value apparently confirms the expected decrease of plunge between Fort Montgomery (55° to 60°) and Highland Falls.

Hell Hole fault.—Hell Hole is the precipitous gorge of Popolopen Creek where it crosses the Storm King granite ridge between Bear Mountain and The Torne. The creek undoubtedly started as a tributary of the Hudson River and is subsequent upon the structures of the Highlands complex. The deeply sculptured stream valley follows the strike of foliation in its lower drowned portion. Thence the gorge becomes narrower, winding, and increasingly rugged. The roughly S-shaped path resulted from stream adjustment on locally prominent longi-

tudinal and cross joints, occasionally controlled by the trend of the rock structure (Pl. 4).

Apparently this type of structural control influenced the relatively straight stream course in Hell Hole very little. Its approximate east-west trend is athwart the flow structure and The Storm King granite underlying these ridges was probably intruded as a single sheet conformable with the synclinal structure of the country rocks. Subsequent erosion of a structural unit of quite uniform resistance would normally produce a series of accordant hill tops.

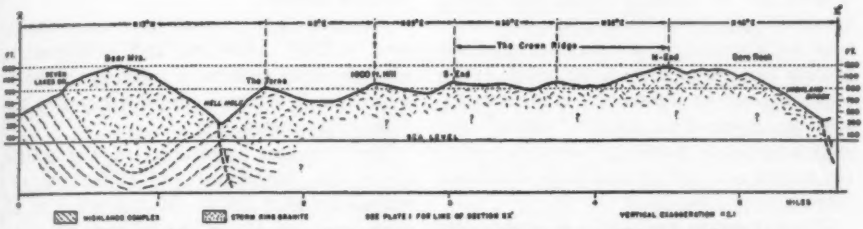


FIGURE 11.—PROFILE AND SECTION FROM BEAR MOUNTAIN TO HIGHLAND BROOK
Curvilinear section (XX′ of Pl. 1) follows topographic crest of Storm King granite intrusive.

prevailing joint orientations in the Storm King granite, and an unusually large basalt dike (50 feet wide) crossing the floor of Hell Hole (07.52–00.47) near the upper end of the gorge does not affect the direction of headward erosion (Pl. 5). This is considered significant, because agents of denudation generally tend to seek basic rocks of this kind in the resistant granite (Thompson, 1936, p. 1837). Thus faulting is the only logical cause of this remarkable transverse valley across half a mile of massive granite.

Despite painstaking investigation, the writer found only one outcrop revealing direct evidence of faulting. A badly weathered, unhealed zone of crushed Storm King granite is exposed in the south wall of upper Hell Hole (07.58–00.43). Its continuation is unfortunately lost under debris from the construction of US 6 which mantles most of the Bear Mountain side of the valley. No similar fracturing is evident elsewhere in the rock floor or walls of the gorge. It is suggested, therefore, that faulting occurred on the south side of Hell Hole. Assuming a steep northerly dip of the fault surface, Popolopen Creek not only excavated the crush material but gradually shifted its course northward—*i.e.*, down-dip.

High-angle faulting with a large vertical component and downthrow to the north is indicated by geomorphic evidence. Figure 11 shows a longitudinal profile following the arcuate crest line from Bear Mountain to Bare Rock. The topographic crest of Figure 11, however, shows that the elevations of the ridge tops drop sharply across Hell Hole from 1305 feet on Bear Mountain to 941 feet on The Torne. From this point northward they rise gradually until the approximate Bear Mountain level is again attained at the north end of The Crown Ridge (1290 feet). The writer suggests that this is the direct result of tilting of the Storm King granite intrusion due to differential block (tear) faulting at Hell Hole and Highland Brook (Pl. 1).

Berkey and Rice (1919, map) did not show this fault but terminated the Storm King granite abruptly along Highland Brook. It is difficult to see how the granite could disappear beneath the Highlands complex along a linear escarpment, some 800 feet high and 1¾ miles long, trending perpendicular to the northeast-plunging structure (Thompson, 1936, Fig. 9, p. 1845) without faulting. Field relations practically demand the presence of a high-angle cross fault analogous to the Hell Hole fault. Downthrow on the northeast is also suggested by the absence of any Storm King granite exposures northeast of the probable fault line in the West Point-Constitution Island-Cold Spring belt (Berkey and Rice, 1919, map).

Colony (1933, p. 27) refers to the "cross fault along Popolopen Creek, on the west side of the Hudson River" as an example of Triassic normal faulting. It is not clear whether he meant the Hell Hole fault or the cross fault in

the valley of upper Popolopen Creek between Cranberry Brook and Forest of Dean (Pl. 1). Although both dislocations could have been the result of the same period of regional deformation, field relations do not permit linking them into a single fault surface.

The writer disagrees with the assumption of Triassic age for either of these cross faults. More probably, block faulting was directly related to Paleozoic thrusts along the Timp Pass-Hudson River and Crown Ridge faults. As the Highlands complex was overthrust to the northwest the Storm King granite "barrier" was under a constantly increasing load which eventually caused tear faulting and block subsidence between such cross faults (Pl. 1).

This mechanism would help to explain two puzzling features at Bear Mountain. The effects of faulting in Hell Hole seem to be confined to the Storm King granite. Neither geomorphic nor structural evidence allows extension of the fault zone far into the country rocks beyond the upper and lower granite contacts. Also the dying out of the Crown Ridge fault at The Torne becomes more plausible, if simultaneous block faulting occurred at Hell Hole to relieve the load strain produced by overthrusting to the northwest.

Petrofabric Analysis

General Considerations

Oriented specimens of Storm King granite from Bear Mountain were collected for petrofabric analysis. The rather clear megascopic structures of the intrusive led to the erroneous belief that this line of investigation would corroborate the field evidence. However, the statistical study of the space-lattice orientation of quartz supports neither the preconceived notions of the writer nor the published findings of Sander and Felkel (1929) and others. Until more statistical data on the structural petrology of the Storm King granite are available, any conclusions are only tentative.

Procedures

Petrofabric investigation was confined to quartz representing ±25 per cent of the rock by volume.

At many places strong alteration and an insufficient number of grains rendered hornblende (±10 per cent by volume) unsuitable for optical orientation and statistical treatment. The definite megascopic orientation of this mineral was substantiated by the attitude of cleavage traces and crystal outlines in the oriented sections.

Feldspars (±60 per cent by volume) are almost entirely microcline and microperthite whose intricate patterns of twinning and intergrowth made accurate optical orientation too difficult to attempt.

The optic-axis orientation of the quartz grains was determined and plotted as described by Haff (1938). Magnification of 25 times was used for all optical measurements. Cedar oil (solvent: benzene) as suggested by Emmons (1943, p. 14) proved superior to glycerine for mounting thin sections and glass hemispheres on the Universal stage. It permitted greater ease of slide movement, an important consideration in petrofabric work, because of its lesser tendency to become "tacky" with time.

Orientation of grains showing slight to moderate wavy extinction was based on median positions of extinction. Occasional grains with mosaic-type of extinction defied measurement.

Preliminary petrofabric study showed no significant variation in statistical orientation of the grains as to size, shape, or textural relation. L. Korn (1933); investigating a biotite granite in central Germany, prepared separate contour diagrams for free quartz and inclosed quartz. She too found that, aside from minor variations, the two sets of fabric patterns were similar and comparable. D. Korn (1929, p. 227) had reached the same conclusion. Consequently very small rounded quartz inclusions in the feldspars of the Storm King granite were not measured. This process of selection, usually frowned upon in this type of investigation, increased the speed and accuracy of operation without impairing the statistical results. Long diameters of quartz grains measured varied from ½ to 10 mm.

Point or scatter diagrams based on lower hemisphere projection were counted out by the concentric-pattern method (Lowe, 1946c).

The English equivalent of the complex Ger-

man terminology of structural petrology employed in this discussion follows the usage of Fairbairn (1937, p. 136, and 1942) and Knopf and Ingerson (1938). Simple definitions of the principal nomenclature may be found in the brief, but lucid treatment of the subject by Billings (1942, p. 332–355).

Field and Petrographic Evidence

In regional geological studies, structural petrology is a corollary to accurate field observations and provides checks for the interpretation based on them (Billings, 1942, p. 335). Therefore, the writer summarizes pertinent megascopic structural features of the Storm King granite.

(1) Contacts of the granite are generally sharp and essentially conformable with the synclinal structure of the invaded crystalline foliates.

(2) Predominantly linear structure of the granite is the result of parallelism of prismatic crystals and aggregates of hornblende, and to a lesser degree of the larger elongated quartz grains.

(3) Its rather constant orientation (40° NE.) is essentially parallel to the trend and plunge of the regional tectonic axis.

(4) With few exceptions additional planar structures appear only close to and essentially parallel to the contacts.

(5) Storm King granite at Bear Mountain and elsewhere in the Hudson Highlands exhibits unusual lithologic uniformity. This is significant in a region of lithologically complex crystallines.

(6) There is no field evidence of *major* tectonic deformation of the Bear Mountain intrusive after its emplacement.

The dominant linear structure is considered indicative of the direction of magmatic flow and thus corresponds to the fabric axis '*a*' (Sander, 1930). There is no field evidence of a '*b*' tectonic fabric axis, even near contacts where linear elements in the planar flow layers retain essentially the same geographic orientation as the linear structure throughout the mass of the granite.

Furthermore, studies with the microscope revealed the following petrographic features:

(1) Sequence of crystallization is hornblende —feldspar and quartz—quartz. The last phase of crystallization generally appears to include quartz grains larger than 2 mm. (long diameter). Mutual inclusions of all three minerals and their more intimate intergrowth near contacts suggest more rapid crystallization probably due to greater temperature gradient.

(2) Hornblende is mostly euhedral, with the crystallographic *c* axis coinciding with the long axis of the mineral grains which defines the observed linear structure. Resorption and alteration effects are common and occasionally extreme, leading to a complete change in composition and the destruction of crystal outlines.

(3) Anhedral to subhedral microcline and microperthite form interlocking patterns without any discernible, dimensionally preferred orientation. Poikilitic texture due to inclosed hornblende and quartz grains is observed occasionally. Clear, unstrained quartz frequently replaces more sodic portions of microperthite and rims many feldspar grains.

(4) Quartz is invariably anhedral. Elongation of grains becomes more prominent with increasing size and appears to be aligned roughly parallel to the hornblende prisms. Boundaries of larger grains in contact with feldspars are more irregular, and many are deeply embayed. Weak to moderate wavy extinction is common, though strained and unstrained grains in adjacent and interlocking positions can be observed. Zones of wavy extinction in the larger grains are typically perpendicular or at high angles to the grain elongation except near contacts with the country rock (Fig. 13). Cataclastic fragmentation of grain borders (mortar structure) is present, but not characteristic. Occasional extreme mosaic extinction is associated with more intense fracturing of all minerals. No constant geometrical relation was noted between orientation of visible fractures, bubble trains, and wavy extinction zones (Fig. 13).

Quartz Diagrams

Figure 12 shows collective diagrams of the poles of *c* axes referred to horizontal projection planes to permit uniform geographical orientation and rapid comparison. The rather coarse texture of the granite necessitated measure-

ments of quartz grains in three to five thin sections per locality. Results obtained from sections inclined to the horizontal (elemental diagrams) were incorporated in the collective diagram by rotation about their strike axis, a procedure permissible owing to the homogeneity of the megascopic rock fabric.

Planar and linear flow structures plotted on the diagrams were determined from field data. The orientation of the less conspicuous elongation of the quartz grains was checked by statistical azimuth counts in oriented thin sections (Haff, 1938, p. 571). Their geographical orientation was determined by graphic solution (Lowe, 1946b). Most grains smaller than 2 mm. across are roughly equidimensional. Elongation of shape as well as preferred linear orientation of elongated grains seems to increase with grain size. About 20 per cent of all elongated grains (2 to 10 mm. in length) showed preferred dimensional orientation parallel to the lineation of the hornblende. This statistical percentage rises to more than 50 per cent for the largest grains (5 to 10 mm.).

Interpretation of Results

Quartz orientation.—Three distinct patterns can be distinguished among the petrofabric diagrams.

(1) D1 (Fig. 12) is believed to be typical of the fabric developed close to the walls of the intrusion. It is characterized by a single, strong maximum (greater than 7 per cent) close to the direction in space of the long dimensional axes of both hornblende (*h*) and large quartz grains (*q*), so that these three elements are essentially parallel. Here the long dimensional quartz axes coincide with the crystallographic *c* axes, and both parallel the long axes (both dimensional and crystallographic) of the hornblende crystals.

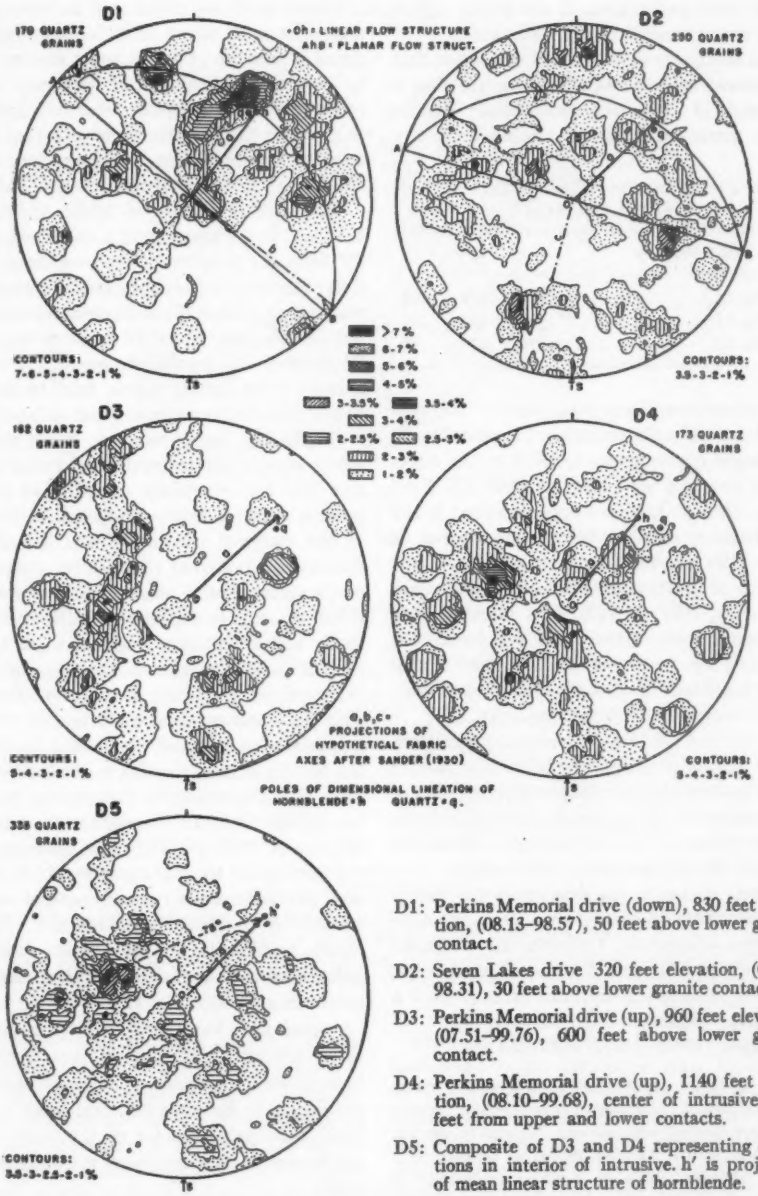
A tendency of secondary maxima to spread out along the '*ab*' fabric plane[8] (planar flow structure) is thought to result from variations in the orientation of flow layers. Field measurements indicate strikes from N. 40° W. to N. 60° W. and dips from 30° to 45° N. for these undulating rather than strictly planar structures. Orientation of linear elements which always lie in the surface of the flow layers varies accordingly. The values given in D1 represent mean orientations.

The secondary girdle arrangement of quartz *c* axes in a small circular arc asymmetrical about the '*a*' axis (in the '*bc*' plane) could be interpreted as the result of plastic flow due to slight continued (tectonic) deformation stresses subsequent to the crystallization of the rock. Under such conditions the deformation plane '*bc*' would be essentially perpendicular to the contact surface. Aside from the difference in the interpretation of the fabric axes (Fairbairn, 1937, p. 112) similar fabric patterns appear in the diagrams of the German investigators (*e.g.*, D. Korn, 1929).

(2) D2 (Fig. 12) represents conditions at a point as close to the same lower contact of the intrusive as that which yielded D1, but ¾ mile farther east. Although both linear and planar flow structures are present and the long dimensional axes of hornblende and large quartz grains are parallel, the space-lattice orientation of quartz is random. The relatively low per cent concentration of the maxima (less than 4 per cent) may be partly due to the greater number of grains (250) measured.

In the light of the reasonably clear orientation pattern obtained from D1, the random distribution of quartz *c* axes in this diagram is more difficult to interpret. A healed zone of intense Storm King granite brecciation starting about 100 feet to the north and extending 800 feet or more into the interior of the intrusive (Pl. 5) suggests a local tectonic disturbance after the emplacement of the granite. (*See* Timp Pass-Hudson River fault.) Petrographic study also reveals strong post-crystallization strains. Fractures are more pronounced and numerous, and many pass through several grains of different minerals. Cataclastic borders of quartz grains are common. Wavy extinction zones in quartz show a greater variation in extinction angles, and many break up into complicated mosaic patterns. Unstrained quartz is rare. Strong limonitization of fractures and interstices points

[8] Fabric axes '*a*', '*b*' and '*c*' of Figure 12 correspond to the orientation scheme of Sander (1930) whereby '*a*' is the "direction of movement" and '*ab*' constitutes an *s*-plane (in this case the planar flow structure); '*b*' and '*c*' are purely hypothetical fulfilling the requirement that the three fabric axes be oriented at right angles to each other.



D1: Perkins Memorial drive (down), 830 feet elevation, (08.13–98.57), 50 feet above lower granite contact.

D2: Seven Lakes drive 320 feet elevation, (08.27–98.31), 30 feet above lower granite contact.

D3: Perkins Memorial drive (up), 960 feet elevation, (07.51–99.76), 600 feet above lower granite contact.

D4: Perkins Memorial drive (up), 1140 feet elevation, (08.10–99.68), center of intrusive, 1500 feet from upper and lower contacts.

D5: Composite of D3 and D4 representing conditions in interior of intrusive. h′ is projection of mean linear structure of hornblende.

FIGURE 12.—COLLECTIVE CONTOUR DIAGRAMS OF QUARTZ *c* AXES IN STORM KING GRANITE

Lower hemisphere projection on Schmidt equal-area net. Horizontal projection planes with uniform north-south orientation.

to the ready penetration of weathering agents along open channels in the rock fabric.

This seemingly supports the explanation that the present random space-lattice orientation is the result of local post-crystallization stresses which destroyed whatever preferred primary orientations may have been present. The preferred dimensional orientation of the hornblende and larger quartz grains parallel to the direction of flow was not visibly affected. (*Cf.* Fairbairn, 1937, p. 112.) No apparent need is felt for assuming regional deformation stresses in view of the generally accepted concept of the mobility of quartz under relatively minor strain (Fairbairn, 1937, p. 112). This orientation of the quartz space lattice could well have been produced by the same stresses (here acting below the elastic limit of the rock fabric) which caused rupture a short distance away.

TABLE 2.—CORRELATION COEFFICIENTS (Pr) OF FABRIC DIAGRAMS

Fabric diagrams	P_r		
D3, Fig. 12	.01	.019	.05
D4, Fig. 12	.01	.022	.05
D5, Fig. 12	.01	.014	.05

(3) D3 and D4 (Fig. 12) represent conditions in the interior of the granite body. Comparable fabric patterns and similar space orientation of the principal maxima indicate essentially identical conditions at the two localities. Composite diagram D5 emphasizes this relationship.

Quartz *c* axes in the interior of the granite mass appear to have only weak to questionable preferred orientation. Therefore the coefficient of correlation (Chayes, 1946) was applied to test whether these fabric diagrams actually show a significant departure from randomness.

Chayes (1946, p. 401) indicated that

> "It is customary to determine r values whose probabilities of being drawn from random populations are, respectively, .05 and .01, and to use the first as indicative and the second as decisive. If a fabric diagram yields an r whose probability is greater than .05 . . . there is no reason to suspect lack of homogeneity; if r has a probability of .01 or less, the existence of a significant departure from randomness in the population is beyond reasonable doubt."

Consequently the patterns of the three fabric diagrams seem to at least indicate inhomogeneity, and the principal point maxima may be considered directions of preferred *c*-axis orientation. The composite diagram (D5) not only shows the best-developed principal maximum, but its coefficient of .014 also approaches the decisive value of .01 most closely (Table 2).

The striking feature of fabric orientation, however, is the presence of a maximum about 75° from the direction of the long dimensional axes of the quartz (and hornblende) grains. This means that, though the dimensional quartz axes, here as along the borders of the intrusive, parallel those of the hornblende grains, the optical axes in these quartz grains tend to be perpendicular to their dimensional elongation.

Quartz was one of the last, if not the last major constituent to crystallize. Probably active flow had essentially ceased. Had viscous laminar flow conditions prevailed as they did in the marginal portions of the intrusive, a statistically preferred dimensional orientation of the quartz grains perpendicular to the flow direction could be expected (Riedel, 1929). Actually such orientation of grains larger than 2 mm. is parallel to the uniform flow direction indicated by the linear hornblende elements. This phenomenon alone could be explained by a special mechanism of progressive freezing of the flowing magma parallel to the encroaching walls thus accounting for the constant parallelism of flow elements throughout the intrusive (Fairbairn, 1937, p. 111). This, however, cannot account for the fact that few, if any, elongated grains have their *c* axes parallel or even subparallel to their long dimension. This is shown graphically by the distinct lack of *c* poles parallel to the '*a*' fabric axis in the composite diagram D5 (Fig. 12).

According to Van Praagh (1947) experiments in the production of synthetic quartz crystals have shown that deposition in the direction perpendicular to the rhombohedral faces is more rapid than perpendicular to the prism faces, and growth in the direction of the *c* axis is most rapid.

This tendency of preferential growth in the direction of the *c* axis would make essentially uniform dimensional and space-lattice orientation the rule rather than the exception.

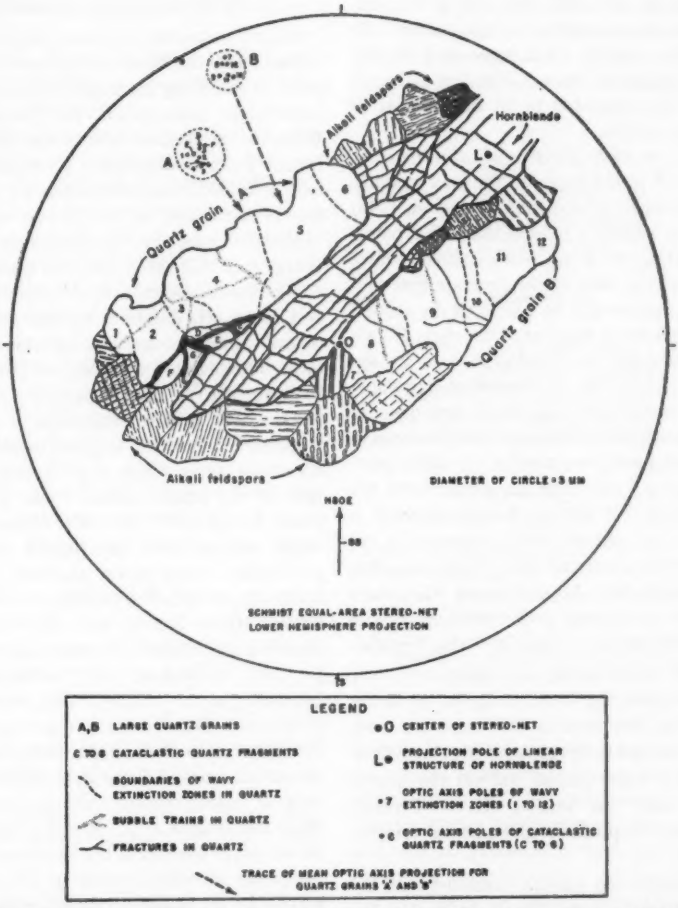

B
A
Alkali feldspars
Hornblende
Quartz grain
Quartz grain B
Alkali feldspars
DIAMETER OF CIRCLE = 5 MM
NADE
SCHMIDT EQUAL-AREA STEREO-NET
LOWER HEMISPHERE PROJECTION
LEGEND
A,B LARGE QUARTZ GRAINS
C TO 6 CATACLASTIC QUARTZ FRAGMENTS
BOUNDARIES OF WAVY EXTINCTION ZONES IN QUARTZ
BUBBLE TRAINS IN QUARTZ
FRACTURES IN QUARTZ
CENTER OF STEREO-NET
PROJECTION POLE OF LINEAR STRUCTURE OF HORNBLENDE
OPTIC AXIS POLES OF WAVY EXTINCTION ZONES (1 TO 12)
OPTIC AXIS POLES OF CATACLASTIC QUARTZ FRAGMENTS (C TO 6)
TRACE OF MEAN OPTIC AXIS PROJECTION FOR QUARTZ GRAINS 'A' AND 'B'

FIGURE 13.—RELATION BETWEEN DIMENSIONAL AND SPACE-LATTICE ORIENTATION OF LARGE QUARTZ GRAINS IN INTERIOR OF STORM KING GRANITE INTRUSIVE

Portion of oriented thin section from locality D4 (Fig. 12) drawn to scale and superposed on plot of quartz optic axes on Schmidt net. Projection plane is plane of thin section. Orientation of sketch and stereo-plot identical.

The Storm King diagrams suggest that conditions near the contact differed from those in the interior of the intrusive. Field evidence indicates that laminar viscous flow prevailed at least near the inclosing walls and parallel to them. The expectable orientation of the linear elements in this environment would be in the direction of flow and parallel to the confining walls as shown by the experiments of Riedel (1929) and discussed by Fairbairn (1937, p. 111) and Knopf and Ingerson (1938, p. 136, 138).

Figure 13 illustrates the characteristic relationships in the interior of Bear Mountain. Wavy extinction of the elongate quartz grains is moderate with relatively minor displacement of *c* axis orientation. Extinction-zone boundaries (zones 1 to 12) are subparallel to the *c* axes of the grains. This feature is characteristic of elongate strained quartz in the interior of the Bear Mountain intrusive. As early as 1896 Mügge recognized a similar phenomenon in a quartz porphyry with extinction angles of the

zones varying up to 20°. Only one of the cataclastic quartz fragments (G) has suffered an appreciable rotation. Fractures and bubble trains in the quartz are not abundant and seem geometrically unrelated to its dimensional or optical orientation.

Comparison with granites of Aberdeenshire, Scotland.—A recent paper on the gray granites of Aberdeenshire in Scotland (Cameron, 1945) gives some valuable information. The petrofabric analysis of mineral orientation (quartz and biotite) was used to demonstrate that preferred orientation can be geometrically related to other structural features of the rock. No attempt was made to explain the data by application of concepts of structural petrology. A rather careful investigation of two quarries in this orthoclase-biotite-muscovite granite gave the following pertinent results: (1) Both quarries are about ½ mile from the contact with the country rock; (2) distinct biotite foliation is interpreted as primary flow structure in accordance with Cloos and Balk; linear structure is also recognized in the larger (Kemnay) quarry; (3) the slightly preferred orientation of quartz from Kemnay suggests "that crystallization proceeded under the influence of some extraneous force and does not appear to be related to the flow structures"; there are more quartz optic axes in the plane of the cross joints than in any other plane; and (4) the quartz diagrams from the smaller (Lower Persley) quarry show haphazard space-lattice orientation.

These results can well be correlated with D3 and D4 (Fig. 12) which represent similar distances from the intrusive contacts and comparable flow features, even though only linear structure is present in the Bear Mountain specimens. The somewhat random orientation quite agrees with the Persley diagrams. If a tendency toward preferred orientation can be argued, a case of agreement with the Kemnay results can be made out. There is at least a lack of quartz axes aligned parallel to the direction of flow and a suggestion of greatest concentration in a plane roughly perpendicular to the flow structure (cross joints) as shown by D5 (Fig. 12).

Differential Solution Hypothesis

General statement.—Absence of field evidence of major tectonic forces acting upon the crystallized Storm King granite is substantiated by microscopic examination. The fracture hypothesis for the reorientation of the quartz space-lattice under metamorphic conditions (Griggs and Bell, 1938) does not explain the anomalous quartz orientation in the interior of the Bear Mountain intrusive. Experimental data by Griggs and Bell agree with the mass statistics on *s*-tectonites assembled by Sander and others. But the quartz diagrams D3 and D4 (Fig. 12) do not exhibit recognizable tectonic fabric patterns. Furthermore there are no indications of the presence of Boehm lamellae (Fairbairn, 1942, p. 10), oriented fractures, or bubble inclusions, which would suggest possible fracture directions along which hypothetical deformation of the quartz lattice could have taken place. Parallelism of the wavy extinction zones might indicate this deformation mechanism, particularly since strain shadows have generally been accepted as products of post-crystallization strain. But the few, admittedly inconclusive experiments of Griggs and Bell (1938, p. 1736) to produce wavy extinction in the laboratory under extremely high pressures gave no evidence that fracturing was responsible for the disturbance of the *c* axis orientation in the extinction zones. A Laue photograph of a fragment of quartz porphyry showing variation in extinction angles of 7° (a value greater than the average observed in the interior of the Bear Mountain intrusive) seems to indicate deformation of the space-lattice by bending of the *c* axis about a bend-axis perpendicular to it and parallel to the unit prism *m*.

Apparently, then, existing hypotheses are inadequate to account for the combination of dimensional and optical orientation of quartz in the interior of the Storm King granite at Bear Mountain. The writer, therefore, tentatively proposes that differential solution under moderate differential stress and at elevated temperatures would explain both the dimensional and petrofabric data obtained (D3, D4, and D5 of Fig. 12).

The hypothesis assumes that:

(1) Quartz crystallized in anhedral grains

without preferred orientation of shape or space-lattice;

(2) During the end stage of crystallization volatile constituents of the magma dissolved parts of the quartz grains;

(3) Those grains which lay with their *c* axes perpendicular to the direction of flow, *i.e.*, parallel to the direction of greatest differential stress, were dissolved most because quartz appears to be most soluble in the general direction of the *c* axis. This would reduce the larger grains that lay in that position to relatively elongated shapes and would destroy the smaller grains.

This hypothesis makes use of three concepts —solubility of quartz in available solvents, differential solubility under differential stress, and increased solubility in the direction of the quartz *c* axis.

Figure 14 shows the relation of probable post-crystallization stresses to the structure of the Storm King granite intrusion at Bear Mountain.

Solubility of quartz.—The solvent action of hydrofluoric acid upon quartz at atmospheric pressure and normal temperature has been used in etching experiments (*e.g.*, Leydolt, 1855). Its solubility in water, however, has been the source of much argument and widely divergent opinions. As recently as 1940 Graton (p. 261) contended that

> "... as yet we are able to count on only truly infinitesimal transport of SiO_2 dissolved in gaseous H_2O."

Ingerson and Morey (1940, p. 779), on the other hand, cite the incidental discovery of 50 g of quartz deposited near the top of a steel bomb from a dismantled steam furnace. The silica was derived from the refractory tubes in the bomb and could not have been carried except through gaseous water. A solubility of $\frac{1}{3}$ of 1 per cent would be indicated under these conditions. Kerr and Armstrong (1943, p. 27) remark that this evidence shows that at temperatures and pressures well above the critical point considerable amounts of silica may be carried by gaseous water.

Kennedy (1944) demonstrated a rapid increase of solubility of silica in water above 200° C. up to 360° C. (critical temperature is 373° C.). At 300 atmospheres pressure quartz solubility increased from .005 per cent at 200° C. to .097 per cent at 360° C. Although a sharp

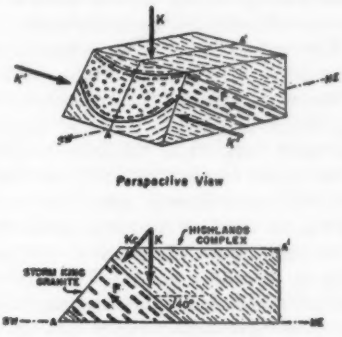


FIGURE 14.—RELATION OF PROBABLE POST-CRYSTALLIZATION STRESSES TO STRUCTURE OF STORM KING GRANITE INTRUSIVE
Idealized schematic block diagram and section.

drop in the solubility curve was noticed above the critical temperature for pressures near the critical pressure (217 atmospheres), exploratory runs at higher pressures indicated an inverse proportional relationship between the specific volume of water vapor and the solubility above 373° C. Thus at 1000 atmospheres pressure, which is equivalent to a rock load of 11,000 feet, the drop in solubility above the critical temperature becomes negligible.

Ingerson and Morey (1940) also touch upon the problem of increased quartz solubility in aqueous solutions containing other chemical constituents, a basic consideration in any attempt to relate experimental evidence to field conditions. Griggs and Bell (1938, p. 1735) used a 10 per cent solution of Na_2CO_3 in experimental deformation of quartz under moderate pressure and temperature, "because of its power of dissolving quartz". At 400° C. and 4000 atmospheres differential pressure the surface of

the quartz test cylinders was etched to a depth of a few tenths of a millimeter in 2 hours.

In view of the inconclusive experimental data on this phase of the quartz solubility problem, evidence bearing on the inverse process of quartz crystallization from solution might be cited. Kerr and Armstrong (1943) discuss an imposing array of experiments in the production of quartz from a variety of compounds containing SiO_2, with H_2O as the major constituent in most cases. Al_2O_3, Na_2O, Fe_2O_3, CaO, K_2O, and other inorganic compounds, when present in the source material, seem to be essential factors in chemical reactions producing quartz as a by-product. The action of such constituents as KCl, LiCl, and $Na_2WO_4 \cdot 2H_2O$ is interpreted as catalytic and especially conducive to the formation of quartz crystals. It could be reasoned that chemical constituents which facilitate the crystallization of quartz would have a similar effect upon its solubility particularly at elevated pressures and temperatures.

Availability of solvents.—In the Storm King granite intrusion, there is abundant field evidence that volatile end-stage constituents of the magma penetrated the crystallized rock either as hot liquids or vapors. Primary joints are commonly filled with epidote, sericite, chlorite, and aplite. Appreciable quantities of fluorite and carbonate are present in the intricate system of veins which have "healed" local brecciation zones in the granite. Rosettes of specular hematite along fractures in this vein material also suggest that the supply of volatiles persisted over an extended period of time.

Differential solution under differential stress.—The elongation of mineral grains has been explained by load crystallization in the presence of a liquid phase according to Riecke's principle (Riecke, 1895). A simple experiment performed by Russell (1935) indicated that "the solubility of a part of a crystal is related to the local stress". Differential solubility at the point of application of localized directed stress on an alum crystal in a chrome alum solution was clearly shown. The change in shape of the crystal was partly due to redeposition of material on the free surfaces where the stress was relatively less. The solution depression produced in Russell's experiment, multiplied manifold, would result in a grain shape with deeply embayed longitudinal borders (Fig. 13).

An objection to the lengthening of the quartz grains in the Storm King granite by recrystallization may be raised in view of the absence of any optical indication of such longitudinal enlargement. But Wright and Hostetter (1917) made optical observations during deformation of sodium chlorate cubes under a nonuniform load of 20 kg/cm², which showed that interference bands crossed boundaries between the original crystal and the newly deposited material. Upon removal of the load the specimens were uniformly isotropic, indicating not only that the new material was deposited in a state of strain, but that the amount of strain was identical (Fairbairn, 1942, p. 55). Thus assuming active elongation of the quartz grains, the effects of this process may well escape detection under the microscope.

The solvents in these experiments functioned under essentially static conditions of pressure and temperature, and initial chemical equilibrium existed between the solid and the liquid phases. A set-up of this type bears little relation to the dynamic environment of geological processes. The material dissolved from the quartz grains by solutions presumably deficient in SiO_2 could well have been transported some distance before physical and chemical conditions were suitable for redeposition in the form of quartz or other silicates. The replacement shells of clear quartz surrounding many feldspar grains might have been so formed.

The factor of differential stress may also have some bearing on the intensity of solution action by varying the local concentration of the solvent. Morey and Fenner (1917, p. 1186) investigated the system H_2O-K_2SiO_3 for the production rather than solution of quartz and found:

> "The primary effect of pressures in systems of this type is to enable the concentration of the water to be kept at the required value. Such an effect of pressure is of great importance, as the concentration of the volatile component is a function of the pressure, and is to be sharply differentiated from the effects due to uniform hydrostatic pressure, which are comparatively small."

Differential solution along crystallographic directions.—Penfield and Meyer (1890) elaborat-

ing on earlier experiments of Leydolt (1855) demonstrated the greatly accelerated rate of solution of quartz in the direction of the crystallographic *c* axis. Etching of a quartz crystal sphere in hydrofluoric acid at room temperature and atmospheric pressure for 3 months produced a remarkable flattening parallel to the *c* axis and comparative elongation in the direction of the *c* axis.

This process was suggested by Griggs and Bell (1938, p. 1743) as possible explanation for the statistical concentration of *c* axes parallel to their hypothetical quartz needles, an essential corollary of the fracture hypothesis of quartz reorientation under tectonic influences.

Comparison with Other Fusion Tectonites

In view of the rather unexpected petrofabric results obtained at Bear Mountain, the limited literature dealing with quartz-lattice orientation in fusion tectonites was studied. Most of the pertinent information appears to have been published in Germany and Austria by members of the "Sander School" of structural petrologists. In all cases preferred dimensional orientation of linear elements appears to be present.

Conclusions drawn from these investigations show a not unexpected agreement with Sander's (1929; 1930; 1934) points of view on fusion tectonites, namely; (1) there is no fundamental difference between the lattice orientations in *B*- or *s*-tectonites and most fusion tectonites, (2) the consequent *c* axis grouping of quartz, therefore, produces maxima and girdles in the '*ac*' or deformation plane perpendicular to the direction of lineation, (3) the orientation of linear elements is parallel to the '*b*' fabric axis, *i.e.*, at right angles to the direction of translation or igneous flow ('*a*' fabric axis).

Close examination of the petrofabric evidence, however, does not always support the rather definite conclusions reached. Johs (1933), for instance, offers eight collective quartz diagrams of which three admittedly show random orientation, while the degree of preferred orientation in the remaining five is questionable. Adequate comparative evaluation of significant maxima and their distribution is greatly hampered by the lack of a "yardstick" applicable to the statistical results of all petrofabric investigations. It is surprising that this well-established method based on mathematical statistics should not have produced a similarly mathematical solution involving the three variables: (1) number of elements measured, (2) numerical percentage concentration of elements, and (3) relative area of concentration in terms of percentage of total diagram area. Joh's five diagrams exhibiting preferred orientation are not readily comparable, since the number of poles per diagram varies from 132 to 847, with attendant variation in per cent concentration of the maxima (greater than 4 per cent and less than 3 per cent, respectively). There is also some doubt as to the degree of significance of the maxima in terms of the areas covered by them.

Fairbairn (1937, p. 112) believed that the discrepancy in interpretation of the linear elements parallel to the '*b*' rather than the '*a*' fabric axis (Johs, 1933; Maroscheck, 1933) is more apparent than real. He suggested that subsequent plastic flow utilizing pre-existing directions and surfaces, without obliterating the primary flow structures, adequately explains the *B*-tectonite fabric patterns obtained. The same interpretation could be applied to the results of D. Korn (1929) and L. Korn (1933).

Gault (1945) interpreted the Pinckneyville quartz diorite of Alabama as a syntectonic intrusion, on the basis of field and petrofabric evidence. Although he assumed the action of regional deformation stresses during and after the intrusive phase, he admitted difficulties concerning the overall interpretation of the quartz diagrams in terms of *B*- or *s*-tectonites. The space-lattice orientation of quartz apparently reflected both the viscous and plastic flow phases of the rock. In any event, the study of the Storm King granite does not permit the assumption of major regional stresses which could have affected it materially.

Summary and Conclusions

The small number of petrofabric analyses available for the present study permits only tentative conclusions. The conspicuous lack of agreement between the quartz orientation in the Storm King granite and the results of comparable studies on similar rocks in Germany and Austria points to many unsolved problems of structural petrology of primary flow rocks.

Although Sander's (1934, p. 47) theoretical classification of fusion tectonites provides for granites "that never took on the same orientation as the enclosing rocks or never took on any orientation in their fabric", none of the factual evidence cited by him appears to be characteristic of this type. The petrofabric results obtained by this writer, and by Cameron (1945) in Scotland, suggest that probably too much emphasis has been placed upon interpreting igneous rock fabrics by the one-sided application of principles established from the study of deformed rocks.

The remarkable mobility of quartz under the influence of stresses of small magnitude and in many different physical environments has been generally recognized (Fairbairn, 1937, p. 112). Quartz orientation in the near-contact zones of the Bear Mountain intrusive supports this contention. Space-lattice orientation in response to viscous laminar flow in one locality seems to have been destroyed by local post-crystallization stresses in another locality representative of the identical geologic environment.

Differential solution of quartz after crystallization is proposed to explain the anomalous relation between dimensional and space-lattice orientation of quartz in the interior of the intrusive. Differential stresses essential for this process are thought to have been supplied by orogenic deformation subsequent to the emplacement of the granite, which was of insufficient magnitude to produce megascopic deformation structures in the rock.

The important role of volatiles in facilitating crystallization, solution, and deformation of quartz has been indicated by numerous laboratory experiments. Few concrete experimental data, however, are as yet available, which could provide valid explanations for these phenomena under the dynamic conditions of geologic processes.

Much needed experimental work along these lines may lead to the acceptance of solution and recrystallization as an important process in the deformation of quartz. It is also evident that little can be accomplished by satisfactory petrofabric analysis of fusion tectonites until mass statistical data from many localities have been compiled. It is to be hoped that structural petrologists will set up valid criteria for distinguishing between the petrofabric phenomena due to primary (flow) and secondary (tectonic) influences.

Origin of the Storm King Granite

Magmatic Intrusion

Until recently few petrologists would have doubted that the Storm King granite was a magmatic intrusion and that it crystallized from a melt of essentially the same composition as the resulting rock. In recent years, however, an increasing number of field investigations have yielded data which suggest that various processes of "granitization" may have formed these igneous-looking rocks in place (Grout, 1941). The writer, too, has presented evidence indicating metasomatic origin of certain granitic rocks associated with the Canada Hill phase of the Highlands complex.

At Bear Mountain and elsewhere in the Hudson Highlands the Storm King granite is an easily recognizable lithologic unit of rather constant mineral composition and texture. Hornblende, the principal dark mineral, evidently crystallized earlier than feldspar and quartz, and its elongated, prismatic crystals are invariably aligned. At Bear Mountain this linear structure has nearly constant spatial orientation parallel to the regional tectonic axis. Feldspar and quartz grains generally do not exhibit a comparable megascopic alignment.

Planar parallelism of minerals is not uncommon near contacts with the Highlands complex where platy structures conform to the orientation of the contact surfaces. Consequently both linear and platy mineral structures are interpreted as primary flow structures. No secondary foliation is in evidence.

The sharp contacts reveal that the Storm King granite and the Highlands complex are essentially concordant. Tabular inclusions of the country rocks were found in the granite only near contacts. They appear to have the same orientation as the near-by wall rocks, and their outlines are sharp and distinct. There is no evidence of fusion or incorporation of these inclusions into the surrounding granite. Extreme proponents of the formation of granites by metasomatism have interpreted such inclusions as "relic structures". Considering the regional features of the Storm King granite the

writer finds that the inclusions can be explained much more plausibly as slivers of country rock separated from the walls by the action of the concordantly intruding granite magma.

Mineralogical changes in the marginal portions of the Storm King granite, giving rise to a contact facies, can be adequately accounted for by reaction of the magma with its enclosing walls (Shand, 1943, p. 92) and are certainly not proof of metasomatism. Furthermore large, irregular masses of coarse pegmatite of unquestionable Storm King origin are common within the granite. Component minerals are unoriented, and borders are characteristically gradational. Such pegmatites were undoubtedly the result of primary segregation in a fluid melt.

Mechanism of Emplacement

The Storm King granite and the Highlands complex are essentially conformable (Pl. 5). The granite occurs as a synclinal sheet (Pl. 1) with a greatly thickened trough (Bear Mountain) and one well-developed limb (The Torne and Crown Ridge). Such an igneous body cannot be called a phacolith, which by definition is confined to the trough or crest of a fold (Harker, 1909, p. 77–78). It might be termed a "synclinal pluton", whereby the shape of the body implies its concordant nature (Billings, 1942, p. 277).

Colony (1921) referred to the Storm King granite as a "quiet" intrusion which appears to have crystallized under no great deformational stress. Petrographic studies by the writer confirm this. Also the lack of preferred lattice orientation of quartz in the granite indicates the absence of marked directional strain during and after crystallization.

In a true phacolith such conditions could prevail, since the space occupied by the intrusion is thought to have been created by folding of the country rock; but in the case of a thick concordant pluton, such as intruded the Bear Mountain syncline, this mechanism is not readily applicable. It is generally thought that intrusions of this type "make room for themselves" by lifting the overlying rocks. The Storm King granite intrusion, which crystallized at depth, must have been under a rather considerable compressive stress, if it was able to lift the certainly great thickness of superincumbent rocks and make room for a pluton 1000 to 3000 feet thick. Why then does the rock fabric show no evidence of having crystallized under great stress? The conclusion seems inevitable that the space occupied by the pluton was made for it rather than by it.

Loewinson-Lessing (1933) suggested a new mechanism to account for the quiet emplacement of the great trap sills of Siberia. He conceived of intrusion in a region of active subsidence where magma was first squeezed up through fractures. Simultaneously the sediments (in this case) sank by plastic deformation into the potential hollow spaces left by the magma, while the latter filled the space thus created by the subsidence of the country rocks. This "exchange of space" would have to be a continuous and interdependent process which could account for the emplacement of large igneous bodies without the necessity of "raising the roof".

The application of this concept to folded regions intruded by concordant plutons is believed to be worthy of consideration. In the Hudson Highlands actually cubic miles of Storm King granite invaded the folded crystallines essentially conformable with their structure. The rather uniform composition of the rock suggests a single magmatic source of probably large size and considerable areal extent.

At Bear Mountain folding of the Highlands complex perhaps provided the original channel for the granite magma in the trough of the syncline, in a manner similar to the formation of a phacolith.

The Storm King granite probably existed as a true magma only in the upper stratigraphic levels of the Highlands terrane. Since it was obviously intruded at some depth below the present surface, the magmatic chamber probably originally occupied space rather close to the folded structures which it later invaded. Thus the weight of the overlying country rocks could have forced them to subside into the potential hollow space previously filled by the now-rising magma. This subsidence was undoubtedly more or less plastic at the depth of the magmatic chamber. But its translation to the higher levels of greater rigidity could have caused actual separation of rock masses along directions of weakness such as bedding, foliation, and fractures. The resulting open spaces, then, could

have been occupied by the liquid magma just as rapidly as they were created. In this way emplacement of large Storm King granite masses in the Hudson Highlands could have been effected without lifting great weights of superincumbent rocks and under relatively low stress conditions.

This mechanism would also help to explain two unusual structural features at Bear Mountain—the sharp downward flexure of the bottom of the syncline, and the great predominance of longitudinal joints in both the granite and the country rocks.

Age Correlation

All that can be said concerning the age of the Storm King granite in the Hudson Highlands is that it is definitely younger than the Canada Hill phase of the Highlands complex. Possibly the Storm King intrusions took place in early Paleozoic time but probably before the Taconic disturbance. The Storm King granite does not appear to be suitable for measuring by the usual radioactive methods.

This granite disappears as a recognizable lithologic unit east of the Ramapo River (Fig. 1), about 12 miles southwest of Bear Mountain. In the Highlands of northern New Jersey and eastern Pennsylvania only mixed granite gneisses have been found. On the basis of chemical composition, mineralogy, and structural similarities the Byram granite gneiss represents the Storm King phase in that region, while the Losee diorite gneiss is probably equivalent to the Canada Hill phase (Berkey and Rice, 1919). Bayley (1941, p. 49) considered the Byram and Losee gneisses essentially of the same approximate age—*i.e.*, Precambrian—since they cross-cut each other in some New Jersey and Pennsylvania localities. The evident age discrepancy of these igneous phases in the two sections of the Highlands may be due to regional structure. The folded rocks in the Highlands plunge consistently northeast. Therefore successively lower stratigraphic horizons are exposed in a southwest direction from the Hudson River. Is it so unlikely, then, that the Byram gneiss represents the earlier, more deep-seated phase of the Storm King intrusion? Although the writer has not examined this rock in the field, it is quite evident from the literature that the Byram gneiss was emplaced in a rather fluid condition possibly even in the form of granitizing solutions. Under the relatively greater pressure and temperature at that depth the magmatic fluid was probably very liquid, mobile and chemically active, since it still contained all its volatile constituents. In the course of its penetration of the upper crustal regions (*i.e.*, the Hudson Highlands) it became more viscous and chemically stable with the loss of volatiles and gradually formed large masses of a melt of quite uniform granitic composition. With increasing viscosity its intrusive action was progressively slowed which accounts for the age discrepancy between the Storm King and Canada Hill granites.

Bayley (1941, p. 48–49), on good evidence, interpreted the Losee and Byram gneisses as differentiates from a single parent magma. Berkey and Rice (1919, p. 33–35) came to the same conclusion regarding the Canada Hill and Storm King granites. The writer can but agree with this concept. The chemical analyses (Table 1) show that the compositions of the two principal granites in the Hudson Highlands differ only in minor respects (*e.g.*, potash-soda ratio). If there were a common, ultimate source, the Canada Hill phase in the Hudson Highlands would represent the most volatile, reactive, and soda-rich components of this parent magma which travelled far ahead of the residual, viscous, and potash-rich melt (Storm King granite).

Conclusions

The existence of a large syncline in the Highlands complex at Bear Mountain furnishes proof of the folded structure of the Hudson Highlands. Precambrian (Grenville?) deformation is believed to have been largely responsible for this structure.

The Storm King granite was intruded as a synclinal pluton concordant with the structure of the earlier crystallines. This is borne out by linear and platy flow structures and conformable contacts of the granite. In addition, rather uniform composition of the granite, sharp contacts, marginal distribution of country rock inclusions, and petrofabric evidence support the concept of quiet magmatic emplacement under relatively little stress. It is suggested that room

for the intrusion was made by the active subsidence of the region into the emptying magmatic chamber while the concurrently rising magma filled the space previously occupied by the country rocks.

Lack of clearly preferred lattice orientation of quartz in the interior of the granite intrusion is believed to be characteristic of post-tectonic igneous rocks. A hypothesis of differntial solution of quartz by end-stage products of magmatic origin is proposed to explain the anomalous relation between dimensional and lattice orientations of the larger quartz grains.

Statistical analysis of the principal joint systems reveals nearly identical orientation patterns in the Highlands complex and the Storm King granite. The predominance of longitudinal joints and the development of cross joints are related to tensional forces acting at different times but in the same direction. The origin of single diagonal joints taking the place of the usual conjugate shear fractures is uncertain. Several tentative explanations are offered.

Interpretation of the major thrust and normal faults at Bear Mountain suggests that both Precambrian and Paleozoic orogenies affected this region and that the results of each can be distinguished to a certain degree.

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LEGEND

(Dark color tints indicate geology based on outcrops, light tints show areas of glacial or soil cover with bedrock geology inferred)

PALEOZOIC(?)

70

Basic dikes, with strike and dip
(Dioritic, diabasic and camptonitic composition

LATE PRE-CAMBRIAN OR EARLY PALEOZOIC (?)

Storm King granite
(Granite east of Hessian Lake and the Playfield mostly contaminated "contact facies")

PRECAMBRIAN, GRENVILLE (?) AND LATER

M

Highlands complex
(Undifferentiated meta-sediments successively affected by the Pochuck diorite and Canada Hill granite phases)
M, *Grenville (?) marble (graphitic, often silicated)*

Contacts, observed and inferred

Faults, observed and inferred

Brecciation zone, with trend

Brecciation zone, trend unknown

Glacial striae, with direction of glacial movement

STRUCTURE

45 (a) | (b) | 30 (c)

40 30 (d) | 20 (e)

Foliation or remnant bedding in the Highlands complex indicating:

(a) Strike and dip;

(b) Strike and vertical dip;

(c) General trend of crumpled folds, with bearing and plunge of fold axes;

(d) Strike and dip, with bearing and plunge of linear structure;

(e) Strike and vertical dip, with plunge of linear structure.

40 (f) | 35 (g)

46 50 (h) | 60 (i)

Structures in the Storm King granite

(f) Planar flow structure, with strike and dip;

(g) Linear flow structure, with bearing and plunge;

(h) Planar and linear flow structures with their respective values;

(i) Contact surface, with strike and dip.

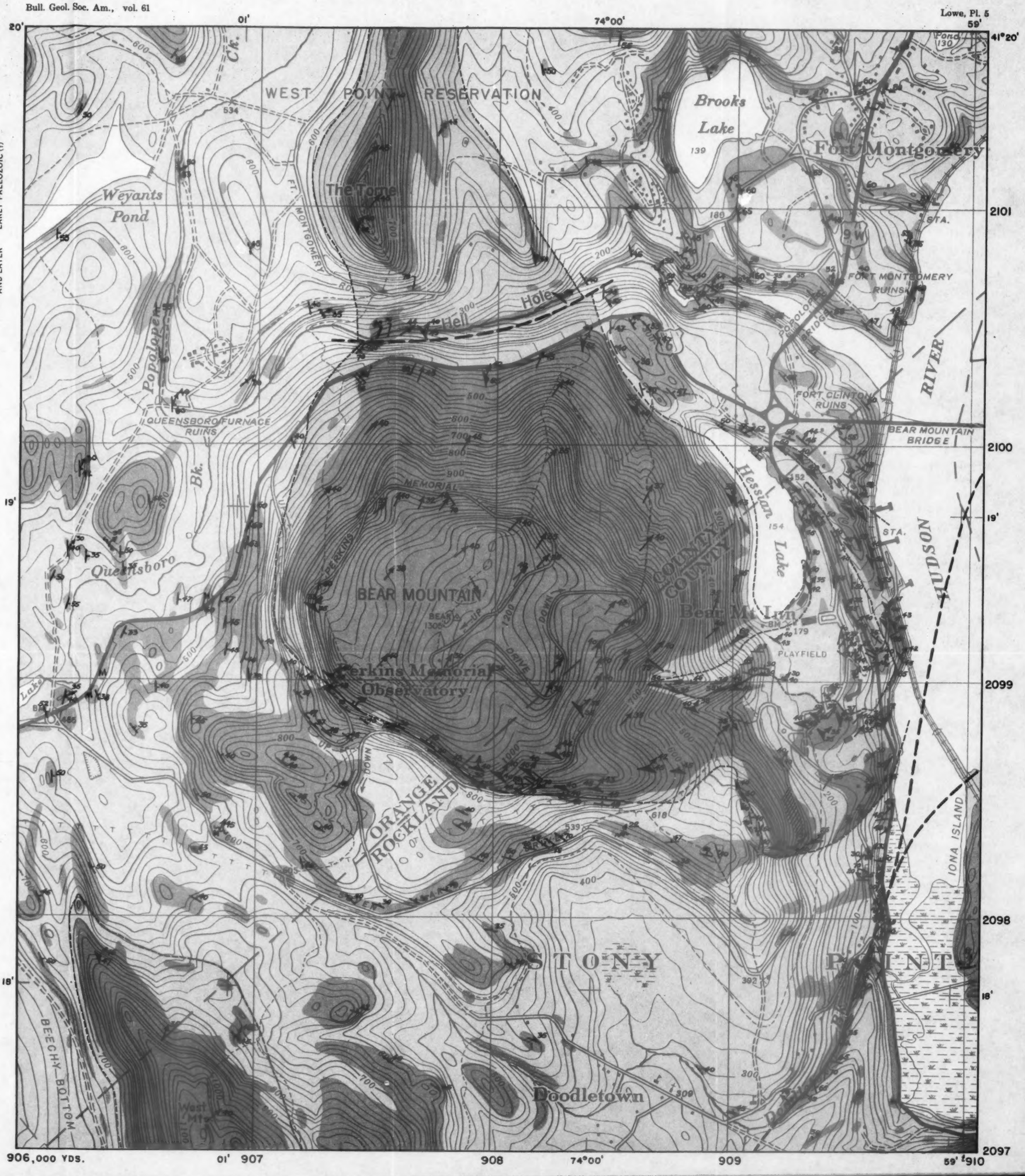
TOPOGRAPHY

U. S. highways *(concrete and macadam)*

Improved roads *(macadam and gravel)*

Dirt roads *(often impassable for cars)*

Trails *(in good condition)*



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(d) (e)

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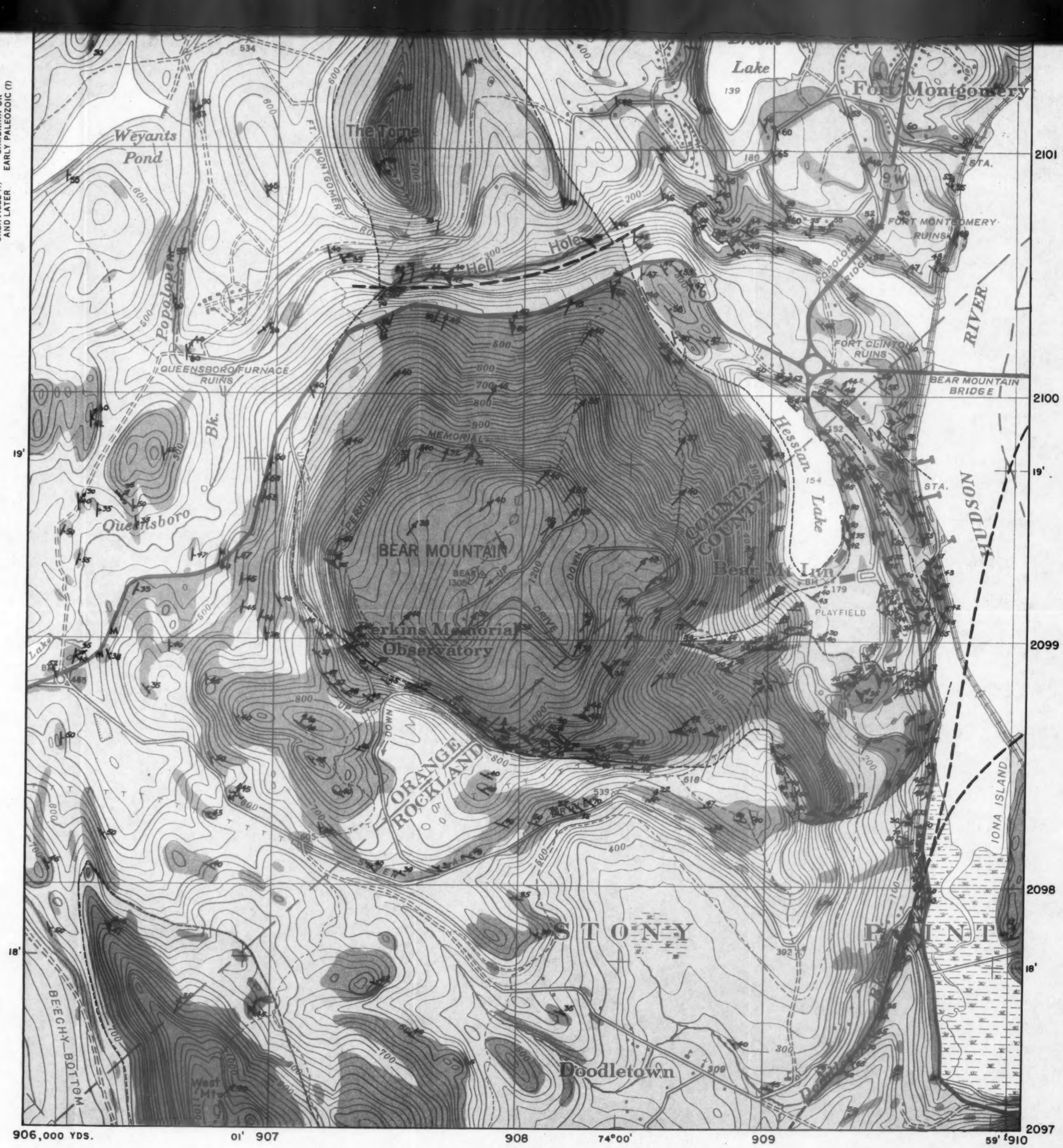

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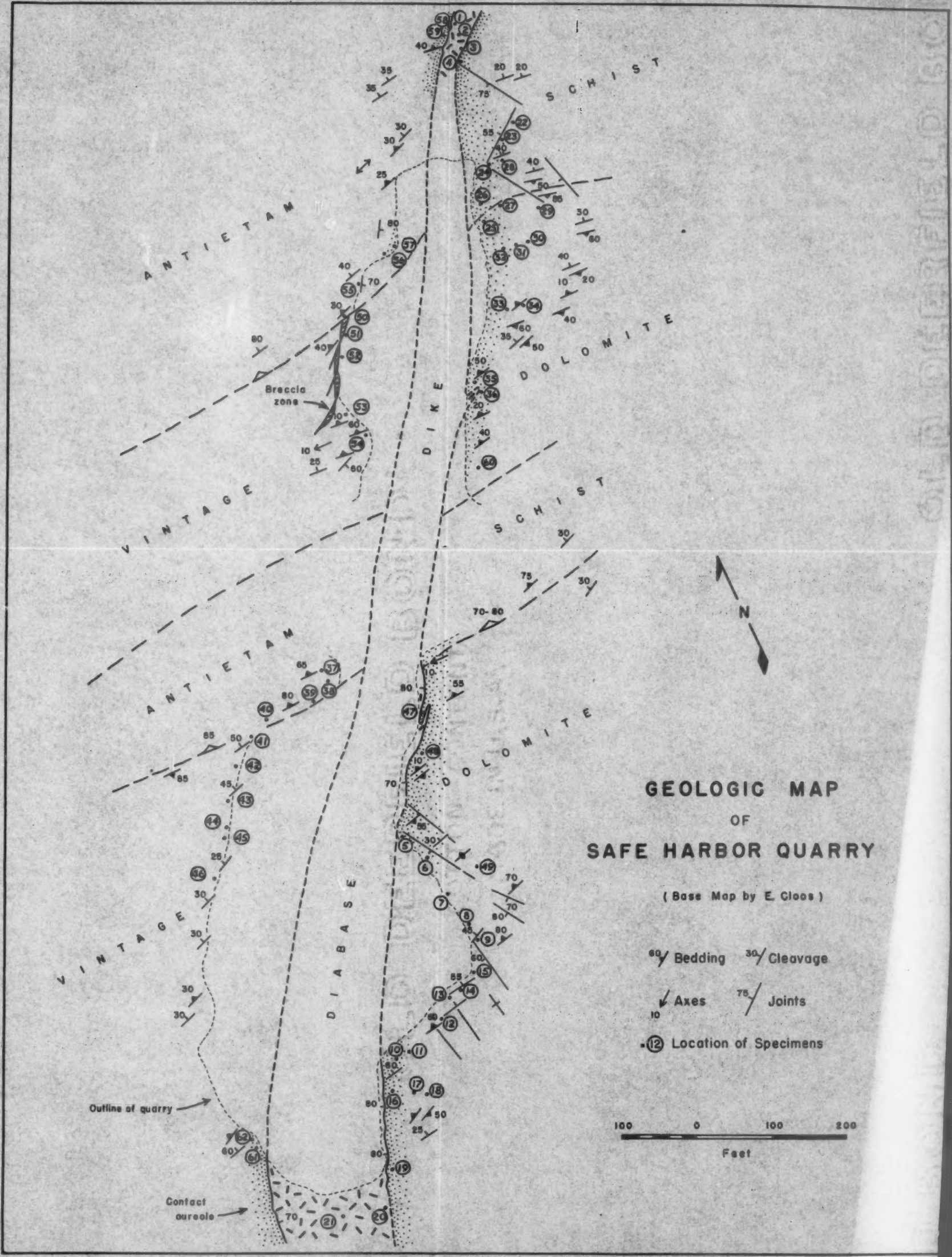
MEAN DECLINATION 11°55' W (1944)

GEOLOGY AND STRUCTURE OF BEAR MOUNTAIN, NEW YORK

Relation of topography to major geologic features

Base map modified from "West Point and Vicinity" sheet, Corps of Engineers, U. S. Army Series, 1:25,000, AMS 4, 1944 (partial color pull).



GEOLOGIC MAP OF SAFE HARBOR QUARRY, PENNSYLVANIA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CONTACT-METAMORPHIC EFFECTS OF TRIASSIC DIABASE AT SAFE HARBOR, PENNSYLVANIA

By Randolph W. Chapman

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Abstract

At Safe Harbor, Pennsylvania, a vertical, northeasterly trending dike of Triassic diabase transects highly folded Antietam schist and Vintage dolomite. A deep quarry in the diabase affords a splendid opportunity to study the contact-metamorphic effects of this intrusion on the country rock, especially along the eastern boundary.

The unaltered Antietam schist is a fine-grained silvery-gray, well-foliated rock containing quartz, feldspar, biotite, and muscovite, but near the intrusion it has been greatly changed. The contact-metamorphic aureole may be divided roughly into two zones: (1) An inner zone, from 5 to 8 feet wide, of dark hornfels containing orthoclase, plagioclase, cordierite, quartz, and chlorite; and (2) an outer zone, of greater and more variable width, of coarser and lighter-colored rock containing mainly plagioclase, quartz, muscovite, biotite, chlorite, and clinozoisite. Heat was the paramount factor in the metamorphism of the inner zone although hydrothermal solutions later played a role. In the outer zone, hydrothermal solutions were chiefly responsible for the alteration.

The unaltered Vintage dolomite is a foliated fine-grained, saccharoid rock composed of dolomite, feldspar, quartz, and phlogopite. The contact aureole, formed by intrusion of the diabase, consists of two parts: (1) an inner zone, some 20 feet wide, where the rock is a dense, black, nonfoliated limestone made up of carbonate (mostly calcite), phlogopite, antigorite (from forsterite), tremolite, diopside, talc, magnetite, and sphalerite; and (2) an outer zone, of variable width, in which the rock is nonfoliated and sugary-textured and contains

carbonate (mostly dolomite), antigorite (from forsterite), phlogopite, and talc (?). The inner zone is the result of strong thermal metamorphism followed by strong hydrothermal alteration. The outer zone was produced by the same two processes operating less intensely.

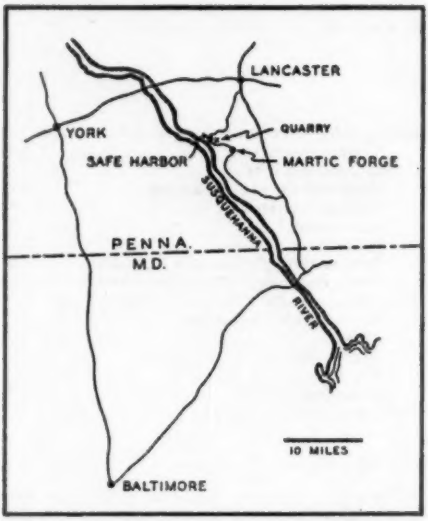

FIGURE 1.—INDEX MAP SHOWING LOCATION OF SAFE HARBOR QUARRY, PENNSYLVANIA

From a consideration of the mineral assemblage of the hornfels close to the diabase, it is apparent that the Antietam schist was thermally metamorphosed under high temperature and moderate pressure of the pyroxene hornfels facies. Although the Vintage dolomite must have suffered similar thermal alteration, its mineral assemblage does not afford conclusive evidence. In the hydrothermal stage, water, iron, and magnesia, and possibly alumina and silica, were added to the schist, and water, magnesia, silica, iron, zinc, and sulphur were brought into the dolomite. However, the magnesia and silica were probably not magmatic in origin.

Judging from published accounts, the Palisades diabase of New Jersey and New York, like the Safe Harbor diabase, metamorphosed its wall rock under conditions of high temperature and moderate pressure characteristic of the pyroxene hornfels facies. The great Whin sill of northern England, however, was cooler and formed contact rocks belonging to the amphibolite facies.

INTRODUCTION

This paper deals with the contact metamorphism of schist and dolomite by a thick dike of Triassic diabase. Safe Harbor quarry, the site of the metamorphism, lies about half a mile due southeast of the village of Safe Harbor in Lancaster County, Pennsylvania, and an equal distance northeast of the Susquehanna River (Fig. 1). It may be reached by road either from Safe Harbor or Martic Forge. Three-fourths of a mile from Safe Harbor, an unimproved road leads northward into the quarry, but since this road is customarily blocked to vehicular traffic, the quarry must be approached on foot.

The quarry is a deep, northeasterly trending pit, 1600 feet long and from 100 to 350 feet wide; the walls, along the east side and near the north end, are nearly 150 feet high (Pl. 2, fig. 1). A small east-west valley divides it transversely into two portions, in this report referred to as the north and south quarries. The excavation was made, over 20 years ago, in the wide Triassic diabase dike which strikes N 30° E through highly folded Cambrian beds of Antietam schist and Vintage dolomite. The diabase was used in the construction of the Safe Harbor dam across the Susquehanna River, and the dolomite has been used since that time for road metal.

As a result of intrusion of the diabase, both the schist and dolomite were metamorphosed within a narrow zone, but the effects produced in these two contrasting rock types were quite different. Deep excavation has exposed the contact aureole and affords an unusually fine opportunity to study these comparative effects.

To the writer's knowledge no systematic petrographic studies have ever been made in this quarry. The general geology of the McCall's Ferry quandrangle, in which Safe Harbor is located, was described by Knopf and Jonas (1929). Stose and Jonas (1939, p. 125) mentioned the southwestward extension of the Safe Harbor diabase dike in their discussion of the petrography and contact-metamorphic effects of the Triassic igneous rocks of York County, Pennsylvania. Cloos and Hietanen (1941) published a revised geological map of the region around Safe Harbor and to the east and a detailed plane-table map of the Safe Harbor quarry which was prepared by Cloos in 1936. A brief article by Tomlinson (1942) described crystals of idiomorphic cordierite from the altered Antietam schist in the quarry and

mentioned other interesting contact phenomena.

Very little information is available on the contact metamorphic effects of the Triassic intrusions of eastern North America. Since the Safe Harbor quarry appeared to offer an exceptional opportunity to obtain data, the writer, in 1947, undertook a detailed study of the rocks to ascertain the character and intensity of alteration caused by the Triassic diabase intrusion. A photostatic copy of the plane-table map prepared by Cloos (Cloos and Hietanen, 1941, fig. 9) was used as a base. A series of systematic traverses were made across the contact zone at various places where it is best exposed, from the dike itself through the altered rock and into the unaltered schist and dolomite. Critical observations were made on the structure, texture, and mineral composition of the rocks encountered, and a large number of specimens were collected from which thin sections were prepared. The optical properties of many minerals were determined, and from these the nature of chemical changes during metamorphism were inferred. The results of this study shed additional light on the contact metamorphic effects of basic hypabyssal intrusions.

Frequent reference is made to specific localities which are designated by numbers on the geological map (Pl. 1). This method of description has the advantage of being both more definite and more intelligible. Then, too, should any reader be particularly interested in this problem, he can visit the specific localities described.

The writer is pleased to acknowledge the assistance of the following in this study. Ernst Cloos called the problem to the writer's attention and kindly permitted the use of his plane-table map of Safe Harbor quarry as a base on which detailed plotting was done. George T. Faust ran X-ray and thermal analyses on the talc, and gave generously of his time and advice in discussing some of the problems encountered in the study. Jewell Glass positively identified crystals of sphalerite by measuring their index of refraction in melts. Leo A. Herrmann visited the quarry with the writer and took the photographs. Gertrude A. Steffe typed the manuscript, and Mildred A. Greenfield and Josephine E. Cleland assisted in proofreading it.

Structural Relationships

Stratified Rocks

Safe Harbor quarry cuts across highly folded Antietam schist and Vintage dolomite, both probably of Cambrian age. Stratigraphically, the Antietam schist, which is the older, grades upward into the Vintage dolomite through a zone roughly 10 feet thick. These rocks have been thrown into a series of tightly compressed folds which strike essentially N 85° E and plunge 15° to 25° SW. The folds are overturned toward the southeast so that their axial planes and most of the beds dip northwest. As a result of folding, the Antietam schist and Vintage dolomite outcrop in the quarry as four parallel alternate bands (Pl. 1).

The high quarry walls, particularly on the west side of the north quarry where they are over 100 feet high, and on the east side of the south quarry, present admirable vertical sections through the folded rocks. Beds range in thickness from a few inches to a foot or more, and are more clearly defined in the Vintage dolomite than in the Antietam schist. Overturning of folds is well exemplified in several places. Locally the bedding is complexly folded (Pl. 2, fig. 3) and at locality 13 a quartz vein, or bed, averaging 0.5 inch thick, has been deformed and pulled out as boudinage.

Schistosity is well developed in both the Antietam schist and Vintage dolomite and for the most part is parallel to the bedding. Both rocks appear to be more heavily jointed on the east side of the quarry than on the west. On Plate 1 the bedding, schistosity, and jointing of the rocks are shown.

The north quarry is bounded on its west side by a huge slickensided wall which represents a fault plane. In general this strikes N 30° E and is essentially vertical. It extends from locality 53 to a point of a few feet southwest of locality 56, cutting directly across the strike of both the Antietam schist and Vintage dolomite. In detail the wall is a smoothly undulating surface. That part which transects the Antietam schist is coated with dark-green serpentine slickensides which indicate movement of the southeast block downward and toward the northeast at an angle of about 30° from the horizontal. Where the plane cuts through the Vintage

dolomite, slickensides of white and pink calcite signify a similar relative displacement.

At locality 55 the schist is closely crumpled and crushed. Between localities 55 and 56 are five large joint planes striking N 55° W and dipping about 85° SW. These extend the full height of the slickensided wall and near the base seem to pass into shear zones.

From localities 51 to 53 is a zone of dolomite breccia that is somewhat discontinuous and locally interrupted by areas of unbrecciated dolomite. The breccia consist of rock fragments, 0.5 inch to 2 inches across, cemented by calcite. Pale-green talc fills some cavities and forms irregular replacements in the breccia fragments.

A zone of intense shattering in the Vintage dolomite was found at locality 6 (Pl. 2, fig. 3). This trends N 20° E and is about vertical and may represent a fault. The heavy mineralization along this zone will be described in the section dealing with the petrography and contact metamorphism of the Vintage dolomite.

The west wall of the north quarry appears to represent either a single fault plane or several closely spaced parallel fault planes antedating the dike. It is believed that this plane or zone of faulting and brecciation and its continuation to the northeast and southwest, or a similar, parallel plane or zone several feet to the east, served as the main channelway along which the diabase dike was intruded. This fault, together with the minor faults and zones of brecciation mentioned above, formed good conduits for the upward and outward movement of fluids from the dike.

Granitic Dikes and Mineral Veins

On both sides of the diabase dike, at the extreme north end of the north quarry, fine pink granitic dikes and coarser pegmatites intrude the Antietam schist. These consist mainly of pink orthoclase and milky quartz. Irregular masses and veins of milky quartz also cut the schist here. Thick quartz veins, some of which are possibly related genetically to the above granite, appear in both the Antietam schist and Vintage dolomite in other parts of the quarry.

At locality 14, in the south quarry, the dolomite is cut by several mineral-bearing veins which parallel the foliation. Some of these are pod-like with a thickness ranging from 1 inch to 8 inches, whereas others are more continuous and have parallel walls 2 to 3 inches apart. The veins consist of large crystals of buff-colored dolomite, some of which are 2 inches across, and small cubes of pyrite set in a matrix of milky quartz. The quartz is younger than the dolomite since thin stringers of it transect rhombohedrons of dolomite. The pyrite is probably contemporaneous with the quartz.

At locality 45 are veins bearing quartz, dolomite, chalcopyrite, pyrite, and tourmaline. Concentrations of pyrite cubes along bedding planes are common in many parts of the quarry, and veins of calcite were observed in several places.

Most of these veins are parallel to original bedding planes although some cut sharply across. They are believed to be generally later than the folding. Since they occur in the schist and dolomite over wide areas outside the quarry, they bear no relation to the intrusion of biabase. Some may be genetically related to the granitic dikes and pegmatites mentioned above.

Diabase Dike and Apophyses

The Triassic diabase dike trends N 30° E and is exposed at intervals for a distance of

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PLATE 2.—SAFE HARBOR QUARRY, PENNSYLVANIA

FIGURE 1. VIEW IN NORTH QUARRY AT SAFE HARBOR, PENNSYLVANIA

Picture taken looking northeast. Diabase dike, now greatly weathered, may be seen on quarry rim just to left of center of picture. Most of the visible country rock is Antietam schist, but the Vintage dolomite appears on the extreme right. Photograph by Leo A. Herrmann.

FIGURE 2. APOPHYSEAL DIABASE DIKE

This small dike is cutting the Vintage dolomite at locality 29 in the north quarry at Safe Harbor, Pennsylvania. Hammer near foot of dike gives scale. Note the columnar jointing. Photograph by Leo A. Herrmann.

FIGURE 3. CONTACT-METAMORPHOSED VINTAGE DOLOMITE

View in south quarry at Safe Harbor, Pennsylvania, showing highly folded and sheared Vintage dolomite near locality 6 now considerably altered to talc and other minerals by hydrothermal solutions from the diabase dike. Photograph by Leo A. Herrmann.

FIGURE 1

FIGURE 2

FIGURE 3

SAFE HARBOR QUARRY, PENNSYLVANIA

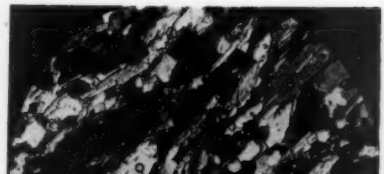

FIGURE 1

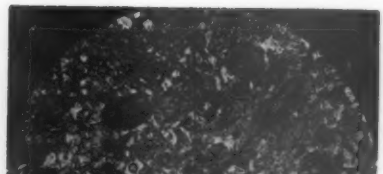

FIGURE 2

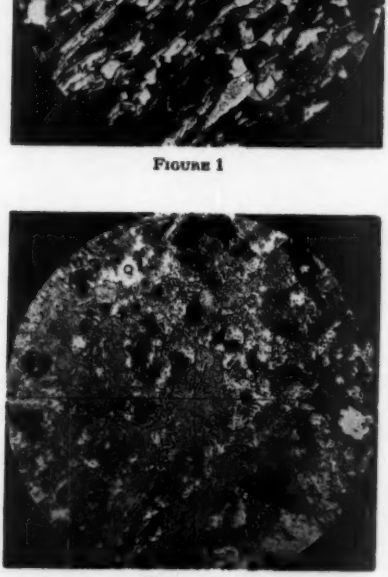

FIGURE 3

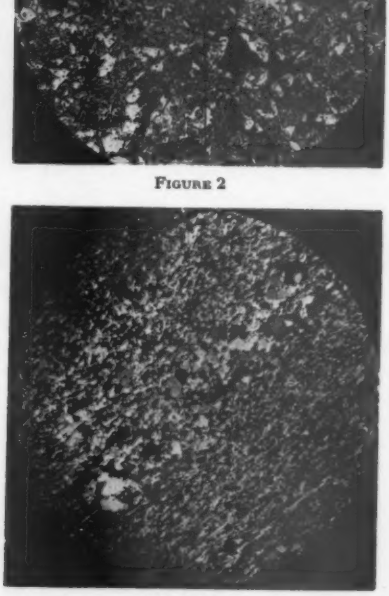

FIGURE 4

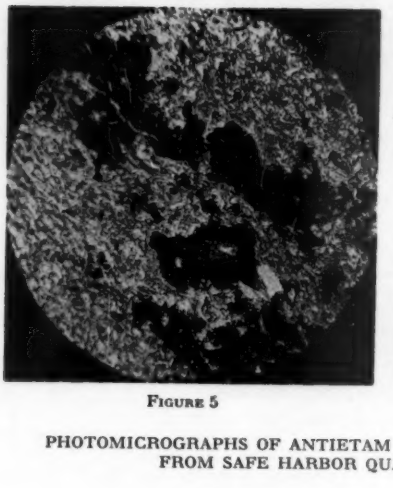

FIGURE 5

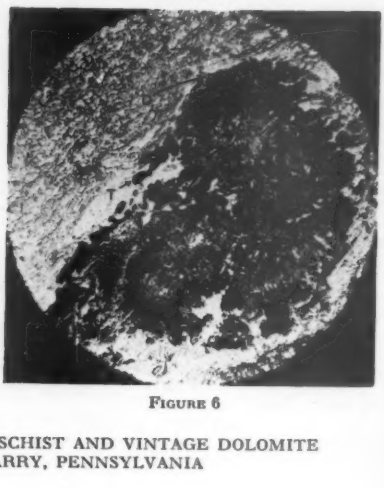

FIGURE 6

PHOTOMICROGRAPHS OF ANTIETAM SCHIST AND VINTAGE DOLOMITE FROM SAFE HARBOR QUARRY, PENNSYLVANIA

about 1600 feet in the quarry. It has been traced about 3 miles to the north of the quarry as well as some distance to the southwest. Plane-table mapping by Cloos (Cloos and Hietanen, 1941, fig. 9) indicates that the width is variable. At the head of the north quarry, it is 35 feet across, but to the south it widens and near the center of the south quarry it reaches a width of probably over 120 feet. The width here is uncertain, however, since the western boundary of the dike has been removed by quarrying. For a reason given later, the writer judges the dike to be thinner at this point than has been indicated by Cloos. At the extreme south, where the dike passes out of the quarry, it has widened to 160 feet.

When Cloos (Cloos and Hietanen, 1941, fig. 9) prepared his plane-table map of the quarry in 1936 he observed the eastern contact between dike and wall rock at several points, but he found that the western boundary had been almost wholly removed by quarrying. The symbols on Plate 1, showing the dips of the contacts, are mainly the result of his observations. In recent years the eastern boundary has been largely obscured by weathering and the slumping of overburden, and the present writer could locate it in only a few places. At locality 4, for example, near the northern end of the north quarry, the eastern contact is well exposed for a short distance and is essentially vertical. A similar observation was made at locality 1 along the western boundary, although here the rock is somewhat weathered. Considered as a whole, the dike appears to be either essentially vertical or dipping very steeply to the northwest, but locally its walls have dips of 70°. Although the dike contacts are invariably sharp, they are not everywhere regular, and small apophyses penetrate the country rock at several localities. These will be discussed later.

In the south quarry, at locality 21, the diabase dike is divided by three persistent sets of joints which strike and dip as follows: N 80° W, 35° SW; N 30° E, 75° NW; and N 30° W,

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PLATE 3.—PHOTOMICROGRAPHS OF SCHIST AND VINTAGE DOLOMITE FROM SAFE HARBOR QUARRY, PENNSYLVANIA

FIGURE 1.—UNALTERED ANTIETAM SCHIST

Composed of biotite (B), muscovite (M), and quartz and albite (clear, somewhat elongate grains, Q). Note good schistosity. Crossed nicols. Diameter of section about 1.5 mm. Specimen from 22 feet southeast of locality 4.

FIGURE 2.—CONTACT-METAMORPHOSED ANTIETAM SCHIST

Shows altered idioblasts of cordierite (C) set in a groundmass of orthoclase and plagioclase (cloudy areas, F), quartz (clear, sharply irregular crystals, Q), and magnetite (small black grains). Note altered pseudohexagonal cordierite twin on crosshair at extreme left. Plane polarized light. Diameter of section about 1.5 mm. Specimen from inner zone of contact aureole adjacent to dike contact at locality 3.

FIGURE 3.—CONTACT-METAMORPHOSED ANTIETAM SCHIST

Consists of large irregular areas of slightly clouded radiating chlorite (C), smaller patches of more densely clouded orthoclase (O) with a little plagioclase, clear aggregates of quartz (Q), and large crystals of specularite (black). Plane polarized light. Diameter of section about 1.5 mm. Specimen from central portion of inner zone of contact aureole 4 feet southeast of locality 4.

FIGURE 4.—UNALTERED VINTAGE DOLOMITE

Rock has good foliation and consists essentially of somewhat elongate crystals of dolomite, together with quartz, plagioclase, and phlogopite. Oriented phlogopite may be seen in lower right portion of section. The thin bed of coarser grains running diagonally through the center is composed mainly of quartz with a little plagioclase. Crossed nicols. Diameter of section about 5 mm. Specimen from locality 9.

FIGURE 5.—CONTACT-METAMORPHOSED VINTAGE DOLOMITE

The lighter part of this section now consists of a confused aggregate of calcite, antigorite, and a little talc. Two crystals of phlogopite (clear areas) are visible near the middle of the southeast portion of the section. The dark bands and areas, parallel to original bedding and schistosity, consist mainly of talc (black), but smaller and clearer inclosures of diopside are visible. The dark nodular body just below and slightly to the right of the center of the picture has a colorless central core of calcite and a little antigorite inclosed by a ring of radiating diopside crystals (gray with high relief), the whole structure being completely surrounded by a broad band of talc (black). Plane polarized light. Diameter of section about 5 mm. Specimen from inner zone of contact aureole at locality 19.

FIGURE 6.—CONTACT-METAMORPHOSED VINTAGE DOLOMITE

This is from the same thin section described in figure 5, and shows a compound, irregular, nodular mass with several cores of fine tremolite fibers (Tr) about which is a regular, more or less continuous reaction rim of diopside granules (D). The whole nodule is surrounded by very discontinuous black patches of talc (T). Light groundmass of rock is chiefly calcite, antigorite, and phlogopite with a little talc. Plane polarized light. Diameter of section about 5 mm. Specimen from inner zone of contact aureole at locality 19.

55° NE. Several other poorer sets were also noted. These separate the rock into rectangular blocks, from 6 inches to 4 feet on a side, and initiate the process of spheroidal weathering. Along the joint faces the rock first alters to a yellow-brown crust which, upon further weathering, finally crumbles to a yellow-brown, granular material. At locality 1, in the north quarry, only one well-defined set of joints exists, the attitude of which is N 10° E, 80° SE. Other irregular joints here assist in dividing the diabase into blocks, 6 inches to 12 inches across, which weather spheroidally and then crumble to fine yellow-brown powder.

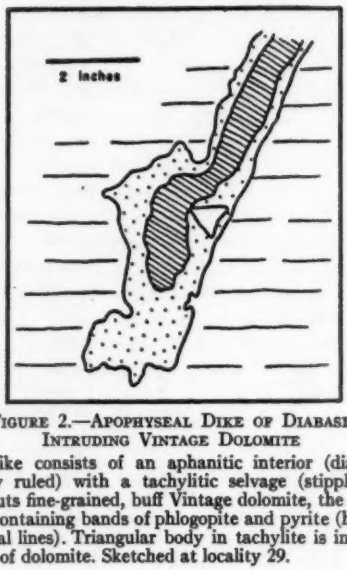

FIGURE 2.—APOPHYSEAL DIKE OF DIABASE INTRUDING VINTAGE DOLOMITE

Dike consists of an aphanitic interior (diagonally ruled) with a tachylitic selvage (stippled). It cuts fine-grained, buff Vintage dolomite, the latter containing bands of phlogopite and pyrite (horizontal lines). Triangular body in tachylite is inclusion of dolomite. Sketched at locality 29.

Some 15 feet southeast of locality 1, the diabase is highly shattered by numerous curved and irregular fractures which may represent a fault zone. Several slickensided surfaces striking N 10° W and dipping vertically are covered with coatings of serpentine and calcite. Thin seams of chrysotile cut through the rock.

The diabase apophyses alluded to above are best exposed from locality 28 to locality 29, the latter point occurring near the east rim of the north quarry. These apophyses, it will be noted, lie from 50 to 80 feet east of the main dike contact. Most of them dip steeply and some are vertical, and thicknesses range from 2 inches to 8 inches. Near locality 28 a large detached block of Antietam schist contains an apophysis, 6 to 7 inches wide near its base, which narrows upward. The walls are smooth and regular and the intrusion shows a tachylitic selvage 6 mm. wide. Columnar joints perpendicular to the walls separate the dike into prisms. Similar small apophyses are more numerous about locality 29 where they cut the Vintage dolomite in various directions (Pl. 2, fig. 2). All are finer-grained near their edges and many show tachylitic borders. Columnar jointing is common. Figure 2 is a sketch of one small apophysis with especially irregular walls.

At locality 47, two somewhat larger satellite dikes, about 3 feet thick and having glassy borders, cut the Vintage dolomite and lie essentially parallel to the main dike contact. One of these shows marked columnar jointing. The other is cut by thin veins of chrysotile. Three similar but smaller apophyses of dense black diabase, 1 to 2 inches thick, intrude the dolomite at locality 16 in the south quarry. At locality 50, on the west side of the north quarry, two fine-grained, green, diabase apophyses, each about 1 inch thick, strike N 10° W and dip 80° SW.

Contact Aureole

Distribution of the contact aureole, insofar as it now crops out in the field, is shown on the geological map (Pl. 1). It is best exposed along the east side of the dike where quarrying operations have been less extensive. At the extreme northern end of the quarry, near locality 4, the aureole is 15 feet across, but near localities 5 and 6 in the south quarry, it measures 65 feet, no doubt because the fault or shear zone here has permitted ready access for solutions. At locality 10 the aureole has narrowed again to about 30 feet. The lack of good outcrops and the removal of rock by quarrying prohibit accurate measurements at other points along the east contact, but it is clear that the width of the contact aureole is variable.

On the west side of the dike the aureole is exposed at only 3 points. At locality 59, at its extreme northern end, it is approximately 15 feet wide. At locality 61, near its southern end, it is at least 8 feet wide and possibly wider.

Altered rock occurs at locality 57, but since the dike contact does not crop out here, the width of the aureole is unknown. Except at these localities, extensive quarrying along the west wall has removed completely the dike boundary and contact aureole. The absence of contact metamorphism along the greater portion of the west quarry wall suggests that the western boundary of the dike actually lies several feet farther east than mapped by Cloos (Cloos and Hietanen, 1941, fig. 9). Cloos' map has been modified by the present writer to that extent.

The boundary between the contact aureole and the unaltered Antietam schist and Vintage dolomite is gradational. This gradation is indicated on the geological map (Pl. 1) by an appropriate pattern.

At locality 52 thin veins of talc fill cavities in the dolomite breccia, but the breccia itself is not metamorphosed. This point probably lies 50 to 60 feet west of the aureole, and the talc has reached here by migrating through the permeable breccia.

Petrography of the Diabase Dike and Apophyses

The diabase is tough and breaks with a conchoidal fracture. At a distance from its contacts it is dark gray and has an ophitic texture which is emphasized by weathering. The principal constituent minerals are less than 1 mm. in diameter. Nearer its boundaries the dike becomes finer, and right at the contacts it is exceedingly dense and black or greenish black. The apophyses, which shoot off from the main dike, commonly have thin glassy selvages 2–6 mm. thick.

In thin section the diabase has a typical ophitic texture with anhedral augite molded between laths of plagioclase. An average mode, in volume per cent, is as follows:

Mineral	*Volume per cent*
Plagioclase	55.0
Augite	42.0
Micrographic intergrowths	1.5
Magnetite	1.0
Apatite, olivine, etc.	0.5
	100.0

The plagioclase forms euhedral laths with diverse orientation. Near the interior of the dike, where the rock is coarser, these laths average 0.6 mm. in length, but toward the margins they become progressively smaller. Normally all crystals show albite twinning and some percliine twinning, but others are untwinned. Zoning is uncommon. Chemical composition of the plagioclase varies slightly from point to point, the average composition being labradorite ($Ab_{40}An_{60}$). Sericitization and some saussuritization have altered many crystals.

Augite consists of anhedral crystals which inclose plagioclase or fill spaces between them. In the coarser specimens, the augite crystals range in cross section from 0.2 to 0.9 mm. and average 0.3 mm. A few grains show twinning. The mineral is colorless and has the optical properties shown in Column 1, Table 1. Many grains are altered along boundaries, cleavages, and fractures to greenish-brown hornblende. Brown biotite is also a common alteration, but whether this comes from the secondary hornblende or directly from the pyroxene is not known. The biotite in turn is changing to chlorite.

Small intergrowths of feldspar and quartz, filling spaces between grains of plagioclase and pyroxene, compose from 0.5 to 2 per cent of the diabase. Some of these are micrographic with tiny wedges of quartz arranged within shapeless masses of feldspar. In others the quartz forms blebs and vermiform patches and the texture becomes granophyric. The feldspar is believed to be sodic plagioclase, although its identity is uncertain due to the intricate relationship with quartz. The interstitial character of these quartz-feldspar intergrowths indicates that they have crystallized late.

At locality 3, the main diabase intrusion is very fine-grained where it contacts the Antietam schist, and it has four zones parallel to the contact. The inner zone, the one adjacent to the normal diabase, ranges in width and consists of a very dense, dark groundmass, apparently of plagioclase and ferromagnesian minerals, but too fine-grained to resolve with certainty. Set in this groundmass are abundant laths of olive-green biotite, averaging 0.26 mm. long, many of which are altering to chlorite. Large oval phenocrysts of augite, 0.2 to 0.9 mm. across,

TABLE 1.—OPTICAL PROPERTIES OF MINERALS FROM SAFE HARBOR QUARRY, PENNSYLVANIA

Optical Properties	1	2	3	4	5	6	7	8	9
Indices of refraction α	—	—	—	—	—	—	1.545	1.608 (?)	1.618
Indices of refraction β	1.684	1.607	1.614	1.657	1.650	1.617	—	1.615	1.621
Indices of refraction γ	—	1.612	1.618	1.657	—	1.617	1.552	1.615	1.621
Optic sign	Positive	Negative	Negative	Negative	Negative	Negative	Negative	Negative	Negative
Optic angle	60°–70°	37°	34°	3°–5°	3°–5°	3°–5°	60°–65°	Very small	1°–2°
Dispersion	r > v, weak	—	—	Medium	r < v, medium	r < v, medium	—	Medium to strong	weak
Pleochroism	None	None	None	X < Y = Z X = pale brown Y = olive brown Z = olive brown	X < Y = Z X = pale brown Y = olive brown Z = olive brown	X < Y = Z X = very pale brown Y = deep brown with olive cast Z = deep brown with olive cast	—	Pale yellow and brown	Pale green and pale yellow green

Optical Properties	10	11	12	13	14	15	16
Indices of refraction α	1.634 (?)	—	—	—	—	—	—
Indices of refraction β	1.637	1.600	1.637	1.635	1.590–2	1.580–90	1.580
Indices of refraction γ	1.637	1.605	1.637	1.635	—	—	—
Optic sign	Negative	Negative	Negative	Negative	Negative	Negative	Negative
Optic angle	1°–2°	33°	3°–5°	3°–5°	ca. 10°	ca. 10°	ca. 10°

Optic sign Optic angle	Negative 1°–2°	Negative 33°	Negative 3°–5°	Negative 3°–5°	Negative ca. 10°	Negative ca. 10°	Negative ca. 10°
Dispersion	weak	—	r < v, weak to medium	r < v, weak to medium	r < v, weak	r < v, weak	r < v, weak
Pleochroism	X < Y = Z X = very pale yellow brown Y = deep green Z deep green	None	X < Y = Z X = very pale brown Y = brown with red cast Z = brown with red cast	X < Y = Z X = very pale brown Y = brown with red cast Z = brown with red cast	X < Y = Z X = colorless Y = yellow brown Z = yellow brown	X < Y = Z X = colorless Y = yellow brown Z = yellow brown	X < Y = Z X = colorless Y = pale yellow brown Z = pale yellow brown

1. Augite from the diabase dike.
2. Muscovite from unaltered Antietam schist 22 feet southeast of locality 4.
3. Muscovite from unaltered Antietam schist at locality 40.
4. Biotite from unaltered Antietam schist 22 feet southeast of locality 4.
5. Biotite from unaltered Antietam schist at locality 40.
6. Biotite from unaltered dolomitic Antietam schist at locality 38.
7. Cordierite from Antietam schist in inner zone of contact aureole (Tomlinson, 1942, p. 648).
8. Chlorite (diabantite), derived from cordierite, from Antietam schist in inner zone of contact aureole.
9. Chlorite from Antietam schist in inner zone of contact aureole, 4 feet east of dike contact.
10. Chlorite (aphrosiderite), derived from biotite, from Antietam schist in outer zone of contact aureole.
11. Muscovite from Antietam schist at locality 59 in outer zone of contact aureole.
12. Biotite from Antietam schist at locality 24 in outer zone of contact aureole.
13. Biotite from Antietam schist at locality 26 in outer zone of contact aureole.
14. Phlogopite from unaltered Vintage dolomite at localities 7 and 45.
15. Phlogopite from Vintage dolomite at locality 6 in inner zone of contact aureole.
16. Phlogopite from Vintage dolomite at locality 19 in inner zone of contact aureole.

are changing to dirty, olive-green biotite (?). Smaller phenocrysts of olivine and plagioclase are common, and rounded crystals of quartz and carbonate appear to be replacing phenocrysts of an earlier mineral, perhaps augite. These early phenocrysts seem to have served as nuclei which are now surrounded by aggregates of biotite. An estimated mode of this zone is shown below:

Mineral	*Volume per cent*
Groundmass	70
Biotite	25
Augite	4
Plagioclase, olivine, carbonate, and quartz	1
	100

In the second zone, which is 1 mm. wide, the dense, dark groundmass mentioned above is predominant and only a few biotite flakes are scattered about. Then comes a thin zone, 0.2 mm. wide, that is considerably lighter than the second zone but of about the same mineral composition. At the outer edge of the dike is a zone 2 mm. wide composed of relatively coarse biotite (50 per cent of rock) and plagioclase and some altered phenocrysts of augite. Here biotite crystals are clustered around the phenocrysts. The abundant development of biotite instead of augite in this border facies may have resulted from the assimilation of alumina and potash from the adjacent schist.

At locality 29 an apophysis of the main dike has developed an interesting tachylitic border. This consists of irregular polygonal cells with glassy exteriors and microcrystalline centers sprinkled with rounded phenocrysts of olivine (?) now altering to antigorite. The tabular microlites in some cells are arranged concentrically, in others they are spherulitic, and in many they have random orientation. The tachylite itself is chiefly yellow-brown and isotropic and its index of refraction is considerably less than balsam. Near its contact with the dolomite it grades into a substance of the same color which is birefringent and has wavy extinction. The indices of refraction of this substance show a wide range and β could be determined only roughly. Approximate values are $\alpha = 1.539$, $\beta = 1.588$ (?), $\gamma = 1.592$. Other optical properties are biaxial negative, dispersion medium, thin fragments are yellow brown, thick fragments orange brown, pleochroism very faint. One grain gave a well-centered acute bisectrix figure with an optic angle of 20°. Several other grains, however, gave acute bisectrices and off-center optic axis figures with an optic angle of over 70°. Interference colors range from nearly zero to first order yellow. These variations in optical properties indicate that the chemical composition is also variable. The substance is believed to be chlorophaeite, an alteration from the tachylite, formed by hydrothermal solutions moving along the dike contact. Peacock and Fuller (1928) describe chlorophaeite from the Columbia River Plateau with indices of refraction ranging from below 1.500 to nearly 1.540 and with a mean refractive index of 1.530.

A thin diabase apophysis at locality 16 has a narrow border of tachylite which in plane polarized light is opaque and shows flow lines parallel to the contact. Small phenocrysts, their outlines suggesting olivine, are rimmed by a fine granular material resembling diopside. The centers of these phenocrysts have been replaced by antigorite and calcite.

As described above, the diabase dike shows a minor amount of alteration consisting of sericitization and saussuritization of plagioclase, replacement of augite by quartz and carbonate, serpentinization of olivine, and the formation of chlorophaeite from tachylite. Most of these changes, which probably took place during the stage of hydrothermal alteration about to be described, involve the addition of hydroxyl and perhaps the concurrent loss of some silica. The latter may have been added to the country rock to assist in the contact mineralization.

Petrography and Contact Metamorphism of the Antietam Schist

Unaltered Schist

The most favorable place to study the unaltered Antietam schist is along the east wall of the north quarry, where the relatively gentle slope of the wall and a broad artificial terrace formed by quarrying permit ready access to both the higher and lower parts of the quarry face (Pl. 2, fig. 1). Along the west wall of the north quarry the Antietam schist is well exposed, but the face is practically vertical, so that its upper portion is rather inaccessible. In

the northwest part of the south quarry, the schist is accessible, but the wall is parallel to the strike of the schistosity.

Outside the contact aureole, the fresh Antietam schist is a dark-gray, silvery rock speckled with flakes of biotite and muscovite. Although it has well-developed schistosity, it is tough and does not split readily. Locally the schist has a vague layering consisting of light quartz-feldspar bands, 2–5 mm. thick, alternating with muscovite bands 0.5 mm. thick. Biotite porphyroblasts are conspicuous, but the diameter of all grains is less than 1 mm. Small crystals of pyrite are localized along bedding planes. Upon weathering, the schist at first becomes lighter gray, apparently because of the bleaching of biotite, and then eventually turns brown.

Where the schist contacts with the Vintage dolomite, the boundary is transitional, the two rocks grading into one another. This transition is well exposed along the foot of the west wall in the north quarry at locality 50, and part way up the east wall near locality 27.

Microscopically the Antietam schist is seen to consist essentially of xenoblastic quartz, feldspar, muscovite, and biotite (Pl. 3, fig. 1). Most of the muscovite and some of the biotite are well oriented, producing a marked schistosity which is further emphasized by elongate crystals of quartz and feldspar. These two latter minerals have an average diameter of 0.1 mm. Some biotite porphyroblasts reach almost 1 mm. and some porphyroblasts of feldspar are 0.33 mm. across. The rock at locality 40 has thin layers of muscovite and layers of even-grained quartz, feldspar, and muscovite alternating with similar layers containing coarse porphyroblasts of feldspar and dolomite.

The mode of the Antietam schist ranges widely as indicated below:

Mineral	*Volume per cent* Range	Average
Feldspar	55–65	60
Quartz	7–15	10
Muscovite	15–18	16
Biotite	6–10	8
Dolomite	3–12	5
Minor constituents	0– 3	1
		100

Dolomite is a minor constituent of the schist except near its contact with the Vintage dolomite where it becomes abundant. Other minor constituents are apatite, zircon, magnetite, pyrite, epidote, and tourmaline.

The feldspar has two modes of occurrence—as small grains in the groundmass and as porphyroblasts. The former, with an average diameter of 0.1 mm., are principally untwinned calcic albite, but one section contains sodic oligoclase and one microcline. Occasional grains show straining, and minute poikiloblastic inclusions are common. Large rounded porphyroblasts of calcic albite, averaging 0.3 mm. in diameter, are abundant in the schist at localities 39 and 40. These abound with poikiloblastic inclusions of biotite, muscovite, quartz, magnetite, and zircon. At least some of these porphyroblasts antedate the foliation of the rock, but others may be younger.

Muscovite forms small oriented grains that for the most part are evenly scattered, although some are concentrated in thin beds. This mineral, from the schist 22 feet southeast of locality 4, has the optical properties shown in Column 2, Table 1, indicating a composition of approximately 50 per cent phengite, 22 per cent muscovite, and 28 per cent Fe'''—muscovite (Winchell, 1933, fig. 197). A specimen from locality 40 has slightly different properties (Column 3, Table 1). The higher indices of this specimen suggest a higher content of iron, the composition of the mineral probably approximating 50 per cent phengite and 50 per cent Fe'''—muscovite (Winchell, 1933, fig. 197).

Brown biotite is coarser than muscovite and is mostly well oriented. Large cross-cutting porphyroblasts are common, nevertheless, and many attain a diameter of 1 mm. Some are skeletal with jagged embayed edges and copious inclusions of quartz, feldspar, magnetite, and zircon. Along the west face of the south quarry between localities 39 and 40, and near locality 50 on the west wall of the north quarry, biotite is altering to chlorite dotted with clusters of epidote crystals.

In the unaltered schist 22 feet southeast of locality 4, the biotite has optical properties as indicated in Column 4, Table 1. These data, applied to Winchell's diagram of the biotite system (Winchell, 1933, fig. 200) indicate a mica with high iron/magnesium ratio, probably siderophyllite. At locality 40 the biotite has slightly lower indices (Column 5, Table 1).

This also is probably siderophyllite with somewhat less iron.

Locality 38 on the west wall of the south quarry is in the transition zone between Antietam schist and Vintage dolomite, and the schist here contains considerable dolomite. Large, irregular biotite porphyroblasts compose 5 per cent of this rock and contain abundant poikiloblastic inclusions of dolomite, feldspar, and quartz. Optical properties of this biotite are shown in Column 6, Table 1. The much lower indices of refraction indicate that this biotite in the transition zone is richer in magnesia with respect to ferrous oxide than the biotite in the normal Antietam schist. Cloos and Hietanen (1941, p. 112) have demonstrated a regional gradation in optical properties between biotite in the Antietam schist and phlogopite in the Vintage dolomite in southeastern Pennsylvania.

Schist of the Contact Aureole

General relations.—The contact aureole in the Antietam schist is confined to the northern part of the north quarry. Near localities 3 and 4 particularly, the contact between dike and schist is revealed, affording an unusual opportunity to study the effects of metamorphism. The aureole here is about 15 feet across. At localities 58 and 59, on the west side of the dike, the aureole has essentially the same width, but the rock is heavily weathered. To the south, near localities 24 and 26, the dike has been removed by quarrying and the quarry floor is now covered by rubble, so that the exact width of contact alteration is unknown, although it appears to be of the order of 60 feet.

The contact aureole in the Antietam schist may be divided into two zones: (1) An inner zone, from 5 to 8 feet wide, throughout which heat was the paramount factor in metamorphism, although hydrothermal solutions later played a role, and (2) an outer zone, of greater and more variable width, where hydrothermal solutions were chiefly responsible for the alteration. The two zones grade into one another and the outer zone is transitional into the unaltered schist outside the aureole. The best complete section through the aureole extends from locality 4 southeastward along the big joint face shown on the geological map (Pl. 1). The description of the inner zone of the contact aureole which follows is based chiefly on data collected along this section but also on information from localities 3 and 58. Description of the outer zone is based on data collected along the above-mentioned section southeast of locality 4, as well as on information from locality 59 to the north and localities 23, 24, 26 and 57 to the south.

Inner zone.—At locality 4 the schist is intruded by pink granite and pegmatite which are older than the diabase dike. At the contact the diabase has chilled against the schist and older pink granite and is very fine-grained. The schist, in a zone about 4 feet wide, has been converted to fine-grained, gray to pink hornfels. The pink color results from abundant orthoclase. Irregular vugs lined with tiny crystals of quartz are numerous. Upon weathering, the altered schist turns yellow brown.

At the inner edge of the inner zone, an inch or two from the dike, all traces of schistosity have disappeared. Microscopically, the rock is a uniform, thoroughly recrystallized, fine-grained aggregate, composed essentially of feldspar, cordierite, quartz, and chlorite (Pl. 3, fig. 2). It is important to note that it contains none of the muscovite or biotite of the original schist. The grain size ranges from 0.08 to 0.3 mm. A rough mode of this rock is shown below:

Mineral	*Volume per cent*
Feldspar	70
Cordierite	18
Quartz	5
Chlorite	5
Minor constituents	2
	100

Feldspar makes up about 70 per cent of the rock, most of which is untwinned, although Carlsbad, albite, and pericline twins were noted. All feldspars are densely clouded with minute inclusions and products of sericitization so that it was found impossible to identify them optically. Consequently, the flat surface of a chip of rock was first etched with hydrofluoric acid and then immersed in a concentrated solution of sodium cobaltinitrite. A bright-yellow stain over much of the surface proved that almost all of the feldspar is orthoclase, perhaps about 5 or 10 per cent being plagioclase.

Cordierite is a significant constituent of the altered schist at this point, where, as evenly

distributed crystals, it forms approximately 18 per cent of the specimen. Although Tomlinson (1942) described fresh cordierite from the Safe Harbor quarry, all crystals noted by the writer are badly altered and the mineral was identified by its characteristic form and twinning. Many crystals have distinct idioblastic outlines, but others form irregular masses dotted with minute inclusions. The idioblasts average 0.2 mm. in diameter and have pseudohexagonal bases and rectangular longitudinal sections. Pseudohexagonal cyclic twinning is evident in several crystals. Tomlinson (1942, p. 648) determined the optical properties listed in Column 7, Table 1. No chemical analysis is available, but these indices are higher than those for most cordierite (Winchell, 1933, p. 307–308), a fact suggesting that considerable ferrous iron is replacing magnesia. Tilley (1924, p. 68–69) concluded that cordierite in the contact aureole of the Comrie area has a ferrous iron/magnesia ratio of at least unity; Fermor (Tilley, 1924, p. 70–71) found in India that shale which had been fused by burning coal beds contains strongly pleochroic iron-cordierite almost devoid of magnesia. Thus the ratio of ferrous iron to magnesia in cordierite may be generally higher than is commonly supposed.

Except for a few minute relics, all cordierite noted by the writer has changed to a faintly pleochroic substance that is rich bright yellow to brownish yellow with an olive cast. This has a flaky, micaceous habit and interference colors ranging from pale yellow gray to bright first order yellow, slightly anomalous. Some of this mineral was isolated from a crushed sample of schist and studied in oil immersions. The optical properties listed in Column 8, Table 1 indicate that the mineral is a type of chlorite, probably diabantite. Similar alteration products have been termed chlorophyllite by other writers (Winchell, 1933, p. 309), and this term seems appropriate here. Locally the chlorophyllite was found grading into pale-green, fibrous chlorite in other parts of the rock, and a thin vein of it was found in one section. Tomlinson (1942, p. 648) noted some alteration of cordierite to pinite at Safe Harbor, but this was not observed for certain by the present writer.

Quartz composes 5 per cent of the hornfels adjacent to the dike contact. Another 5 per cent is green chlorite which makes fan-shaped clumps of radiating fibres. This mineral is similar to the chlorite in the altered schist farther from the dike contact which will be described later. Prominent minor constituents include magnetite, zircon, titanite, and pyrite, the latter altering to hematite.

Four feet southeast of the contact, near the middle part of the inner zone, the altered schist is fine-grained and dark greenish gray from abundant chlorite. Some layers have a pinkish cast. Crude schistosity is evident, but the rock is tough and breaks irregularly. Microscopically this rock is seen to be completely recrystallized with only faint traces of foliation (Pl. 3, fig. 3). It consists essentially of aggregates and crude bands of quartz and unaltered feldspar together with irregular masses of chlorite, altered cordierite, altered feldspar, hematite, pyrite, magnetite, and zircon (?). In one slide the rock is divided into rudely polygonal areas with interiors of chlorite, altered cordierite, altered feldspar, and hematite enclosed by rings of quartz and feldspar granules, an arrangement apparently resulting from metamorphic diffusion.

Quartz and feldspar grains in the aggregates and bands are sutured and average 0.1 mm. in diameter. The quartz composes about 25 per cent of the rock, and the feldspar, chiefly unaltered albite (Ab_{90}), makes up 5 per cent.

A different feldspar, that is highly sericitized and perhaps saussuritized, composes nearly 30 per cent of the altered schist here. Its pink color and general appearance in thin section suggest orthoclase.

Two varieties of chlorite are present near the middle part of the inner zone (4 feet southeast of the dike contact) and together compose about 25 per cent. The predominant type forms large, shapeless, interstitial masses, the distribution of which suggests introduction from without. Each mass is an aggregate of smaller fan-shaped clumps of radiating fibres showing interference colors of ultra blue and rich purple. The optical properties of this mineral, shown in Column 9, Table 1, indicate that it is also diabantite but closer to the daphnite-amesite side of the diagram (Winchell, 1933, fig. 203), i.e., richer in alumina, than is the chlorite that is altering from cordierite.

The other type of chlorite, pseudomorphous after biotite, is deeper green and constitutes

rectangular crystals studded with fine magnetite granules and somewhat larger grains of epidote. It is more strongly pleochroic than the previous type and its interference colors are especially deep purple, suggesting a higher iron content.

About 10 per cent of the rock consists of dark hazy areas made up chiefly of chlorite, dusted with extremely fine-grained magnetite and possibly biotite and epidote. Many of these areas are irregular, some elongate, and others rounded. Some rectangular ones are fresh enough to show faint birefringence and parallel extinction; they are believed to be altered incipient crystals of cordierite.

Granular aggregates and fan-shaped, needle-like growths of specularite are scattered at random and compose 5 per cent of the altered schist near the middle part of the inner zone. Closely associated with the hematite is a dirty, olive-green biotite with notable pleochroism and birefringence. In a number of places this mineral can be traced into the pale-green chlorite described above.

Outer zone.—It has been pointed out that the inner zone of the contact aureole grades into the outer zone, and on the east side of the quarry the transition takes place between 5 and 8 feet from the dike contact. Twelve feet southeast of locality 4, within the outer zone of contact metamorphism, the altered schist is considerably coarser-grained and much lighter in color than the rock near the dike contact. Its definite schistosity and silvery luster are similar to those of the typical unaltered schist farther east. However, chlorite has replaced biotite porphyroblasts, producing a greenish cast. Weathering turns the schist yellow-brown.

Microscopically, the plagioclase and quartz in this rock show sutured boundaries, many constituents have a dirty, altered appearance, masses of fibrous, radiating chlorite are present, and porphyroblasts of biotite are altered to chlorite. The mode has been determined as follows:

Mineral	*Volume per cent*
Plagioclase	60
Quartz	15
Muscovite	13
Chlorite	11
Minor constituents	1
	100

Minor constituents include apatite, magnetite, pyrite, hematite, and epidote.

The plagioclase, which is albite-oligoclase (Ab_{80}), is mostly xenoblastic. Many crystals are clouded with fine sericite, and poikiloblastic inclusions of quartz, muscovite, and magnetite are abundant. Much of the muscovite is dusted by either fine alteration products or by minute inclusions.

The large cross-cutting porphyroblasts of biotite, so characteristic of the unaltered Antietam schist, have been completely changed here to chlorite, which is strongly pleochroic in deep green and pale yellow brown. The chlorite is peppered with minute crystals of epidote and some of magnetite. Column 10, Table 1, shows the optical properties as determined in oil immersions. Winchell's (1933, fig. 203) diagram indicates that this mineral is aphrosiderite, a chlorite somewhat richer in iron and aluminum than either of the two chlorites already described.

Southeastward the altered rock of the outer zone of contact metamorphism passes imperceptibly into unaltered Antietam schist.

On the western side of the dike in the extreme northern part of the north quarry, the contact aureole is exposed but the rock is deeply weathered. A relatively fresh specimen was collected from locality 59, however, 10 feet northwest of the contact. This rock resembles closely that described above which was collected on the east side of the dike, 12 feet southeast of the contact. It was originally typical Antietam schist, consisting of plagioclase, quartz, biotite, and muscovite, but much of its foliation was destroyed by recrystallization. Some quartz-rich and mica-rich layers are still evident, however, and some quartz crystals appear to have been sliced off abruptly perhaps by shearing of the rock. Muscovite forms wavy bands, and the large flakes appear to have been dissolved and recrystallized along their edges to a felty mat of sericite. Plagioclase is highly sericitized, and large porphyroblasts of brown biotite have altered to chlorite with the separation of magnetite and epidote. Several large crystals of epidote are present, and green radiating chlorite forms interstitial masses between quartz and plagioclase grains.

Column 11, Table 1, shows the optical properties of the muscovite from locality 59. The chemical composition, deduced from Winchell's

(1933, fig. 197) diagram, is 70 per cent phengite, 10 per cent muscovite, and 20 per cent Fe'''—muscovite. This mineral is apparently richer in magnesia than the muscovite from the unaltered schist 22 feet southeast of locality 4.

The positions of localities 23, 24, and 26 in the metamorphic aureole are not known precisely since the dike contact is not exposed in their vicinity. Texturally and mineralogically, however, these rocks belong to the outer zone of contact metamorphism. At localities 23 and 24 the altered schist resembles hornfels and is a heavy, massive, nonfoliated rock that breaks conchoidally. Colors range from greenish gray to nearly black with a pink cast. The constituent minerals are mostly less than 1 mm. across, although porphyroblasts of biotite reach a diameter of 1 mm.

Microscopically, this rock has the recrystallized texture common to the other specimens of schist in the outer zone of the aureole, but contains clinozoisite and an unusual amount of dirty-brown biotite set in a background of quartz and altered feldspar. An estimated mode of this rock is shown below:

Mineral	*Volume per cent*
Feldspar	66
Quartz	10
Biotite	15
Muscovite	3
Clinozoisite	3
Chlorite-epidote aggregates	2
Minor constituents	1
	100

Minor constituents are carbonate, apatite, tourmaline, and magnetite.

The feldspar is chiefly untwinned albite-oligoclase, although possible orthoclase was noted. Sericitization of crystals is common.

Most crystals of biotite are well oriented and are stringy and dirty with minute inclusions and alterations. Others are porphyroblastic. Many crystals of both types are bleached to nearly colorless mica with the separation of granules of magnetite along cleavage cracks and crystal boundaries. Considerable biotite has altered to pale-green fibrous chlorite, tiny crystals of epidote, or to confused aggregates of both of these minerals, these alterations having been accompanied by the separation of magnetite. Complete transitions were noted between fresh biotite and chlorite-epidote aggregates.

Biotite from the metamorphosed schist at locality 24 has the optical properties shown in Column 12, Table 1. These indicate a lower iron-magnesium ratio (Winchell, 1933, fig. 200) than for the mica from the unaltered schist, so the mineral is probably closer to eastonite.

At locality 26 the rock is light gray green, more schistose than at localities 23 or 24, and contains many cross-cutting biotites. An estimated mode follows:

Mineral	*Volume per cent*
Feldspar	45
Quartz	10
Biotite	10
Clinozoisite	30
Muscovite	4
Minor constituents	1
	100

Minor constituents are calcite, apatite, zircon, magnetite, and pyrite.

The feldspar is chiefly albite-oligoclase (Ab_{90}), some grains of which show albite twinning. Several constituents were questionably identified as orthoclase. Since the untwinned plagioclase resembles quartz closely, the percentages shown in the mode are not certain. The biotite is optically similar to that at locality 24 (Column 13, Table 1) and is altered to pale yellow-green chlorite with a fibrous radiating habit. Muscovite flakes have subparallel arrangement.

The above minerals are generously interspersed with aggregates of clinozoisite, the abundance and distribution of which suggest that it was produced principally at the expense of plagioclase. Many plagioclase crystals, particularly along boundaries, appear to have altered to clinozoisite.

At locality 57, across the quarry, the rock resembles that at locality 26. Irregular confused masses, that appear to consist chiefly of clinozoisite, are replacing plagioclase.

Contact Metamorphism of the Schist

The typical Antietam schist at Safe Harbor is fine-grained, well-foliated, and consists mainly of calcic albite, quartz, muscovite, and biotite, but where it grades into the Vintage dolomite it contains some carbonate. It was into

this rock that the diabase dike was injected, altering the schist around it. The contact-metamorphic history may be divided into two stages: (1) A thermal stage resulting from the heating of the schist by the hot diabase ("contact" pyrometamorphism of the economic geologist), and (2) a hydrothermal stage caused by hot solutions rising from depth along the dike contact and along fissures and faults ("contact" pyrometasomatism of the economic geologist). The thermal stage is represented mainly by the rocks of the inner zone of the aureole. The amount of heat in an intrusion of the size of the Safe Harbor dike is, of course, small, but its transmission through the country rock probably resulted from conduction and from the movement along fractures of hot igneous gases which came from the cooling dike and from below. The thermal zone is narrow, attaining a maximum width of from 5 to 8 feet so far as can be determined. The later hydrothermal stage not only affected those rocks that had already been thermally altered, but extended beyond, forming an outer zone that shows mainly hydrothermal changes.

Probably the most conspicuous effect of the rising temperature was complete recrystallization of the fine-grained schist to a compact hornfelsic mosaic, with the destruction of all schistosity. Quartz and sodic plagioclase were recrystallized side by side, the latter possibly becoming richer in lime although this cannot be definitely established. With the loss of water from the highly heated schist, muscovite became unstable and was converted to orthoclase in the inner zone of the aureole. However, as already noted, muscovite is still present in the hydrothermally altered zone farther from the intrusion. Adjacent to the contact, all biotite likewise disappeared, some of its potash probably going into orthoclase and its other ingredients into cordierite. The reaction between quartz and the micas to form orthoclase and cordierite with rising temperature has long been recognized and may be illustrated by the following equation:

$$6\,KAl_3Si_3O_{10}(OH)_2 + 2\,KMg_3AlSi_3O_{10}(OH)_2 + 15\,SiO_2 \rightleftharpoons 8\,KAlSi_3O_8 + 3\,Mg_2Al_4Si_5O_{18}\cdot H_2O + 5\,H_2O$$

(muscovite) (biotite) (quartz) (orthoclase) (cordierite)

Farther from the contact, but still within the inner zone of alteration, some biotite remained as coarse porphyroblasts, now altered to chlorite. This probably means that complete equilibrium was not attained in this part of the aureole during the brief period when the temperature was at its peak.

The formation of cordierite was a significant mineral change in the thermal zone of alteration. Near the intrusion, idioblastic crystals formed in a groundmass of orthoclase, plagioclase, and quartz, but farther from the intrusion, irregular, shapeless masses, dotted with minute poikiloblastic inclusions, were developed. Since cordierite is a mineral with weak crystallizing power, it is often unable to rid itself of the groundmass material in which it grows. In the writer's judgment the ingredients of the cordierite were probably derived mainly from the biotite of the unaltered schist, the distribution of the cordierite suggesting that each crystal formed around a nuclear core of biotite flakes.

Heating of the schist seems to have induced some recrystallization in the outer zone of the aureole but no mineral changes were brought about there.

The thermal stage of metamorphism was followed by a hydrothermal stage during which hot aqueous fluids arose along the dike contacts and spread laterally into the country rock. In this report these fluids are referred to as "hydrothermal solutions" although it is possible that some of them may have been gaseous. It is believed that they came from considerable depth rather than directly from the dike itself since the latter appears to have been crystallized at the time and to have been altered by them. Their lateral movement in the country rock was greatly facilitated by bedding and schistosity planes which are nearly normal to the dike contact, and by joints and faults which are abundant. This wave of hydrothermal alteration not only superimposed itself upon the already thermally metamorphosed rocks of the inner zone but moved beyond, developing an outer zone of alteration. Local differences in permeability of the rock probably account for differences in width of this outer zone, an abundance of bedding and schistosity planes, joints, and faults being the predominant factors. Although the incoming solutions un-

doubtedly brought some heat, their major function was the introduction of certain elements from without and the redistribution and recombination of elements already present in the rock.

Texturally, the rocks were little affected by the hydrothermal solutions, the chief changes being mineralogical, both in the thermal zone and in the immediate area surrounding it. In the thermal zone, cordierite became unstable and was converted to yellow chlorite (chlorophyllite). Tomlinson (1942, p. 648) states that some cordierite here has gone to pinite, but this alteration was not detected for certain in any slides studied by the writer. Both of these alterations must have involved the addition of water to the rock and possibly other chemical changes. With the formation of chlorophyllite, for example, there may have been an addition of iron and/or magnesium to the cordierite, and changes in the iron/magnesium ratio, but neither of these could be determined microscopically and there is too little chlorophyllite present to separate for chemical analysis. The change to pinite would undoubtedly mean an addition of potash and water to the original cordierite. Probably all these changes, except that involving water, were accomplished by a recombination of elements already present in the thermal zone.

Another notable effect of the hydrothermal solutions, both on the thermally metamorphosed zone and on the rock some distance beyond, was the alternation of biotite porphyroblasts to chlorite. Most of the biotite in the schist near the diabase was converted to cordierite by thermal metamorphism, but in the outer part of the thermal zone some survived. This is now represented by rectangular crystals of green, strongly pleochroic chlorite dotted with fine grains of magnetite and epidote. Similarly, in the outer zone of the aureole all biotite became unstable under hydrothermal conditions and was converted to chlorite. The outer limit of this biotite alteration marks the outer boundary of the contact aureole. Cloos and Hietanen (1941, p. 136 and pl. 18) state that the biotite in much of the Antietam schist elsewhere shows some alteration to chlorite and this has been noted by the present writer. Regardless of the cause elsewhere, the change in the Safe Harbor quarry is clearly due mainly to hydrothermal solutions from the diabase. This alteration involved the addition of water to the biotite and a recombination of elements already present, with the separation of iron to form magnetite and of calcium, iron, and aluminum to form epidote.

Clinozoisite is found only in the outer zone of the aureole, principally at localities 26 and 57 in the north quarry. Chiefly, it was formed at the expense of sodic plagioclase in the schist under the impetus of hydrothermal solutions. Both localities, 26 and 57, lie in the transition zone between Antietam schist and Vintage dolomite where carbonate is abundant, and it appears that this mineral furnished some of the necessary lime. It is unnecessary to postulate the introduction into the schist of any material to form the clinozoisite.

The generation of large quantities of pale-green chlorite in both the inner and outer zones of the aureole is, perhaps, the most striking effect of the hydrothermal solutions. This mineral (diabantite), poorer in iron and aluminum than the chlorite of the porphyroblasts, makes up as much as 20 per cent of the rock of the inner zone, and in the outer zone occasionally as much as 8 or 10 per cent. Since the unaltered schist clearly contains insufficient iron and magnesium to have formed all of this mineral and the other two types of chlorite mentioned above, we are forced to conclude that the iron and magnesium, together with abundant water, were brought into the contact zone from without. The magnesium, t be sure, was not necessarily igneous in origin and may have been derived from dolomite at depth. Most of the necessary alumina and silica were probably derived from feldspars or micas, although it is possible that some may have been brought in. Without chemical analyses of the altered and unaltered schists, this point cannot be established.

More obvious evidence of the introduction of iron is shown by large crystals of specularite which in one slide, from the inner zone, compose 5 per cent of the specimen. Lesser amounts and smaller crystals are found in the outer zone. Both the abundance and distribution of this mineral justify the conclusion that iron was brought in from the diabase by moving solutions.

Summarily, the earlier stage of metamorphism was a thermal alteration of the schist in

a zone 5 to 8 feet wide. The elevated temperature resulted, especially near the contact, in the complete recrystallization of the schist, the disappearance of all muscovite and most biotite, and the generation of orthoclase and cordierite. This stage was followed by the appearance of hot aqueous solutions along the dike contact which swept outward through the thermal zone and into the unaltered schist. These certainly brought iron and magnesia in abundance and possibly alumina and silica, although the latter two may have been derived from original feldspars and micas. In the thermal zone, cordierite was changed to chlorophyllite, the remaining biotite was converted to chlorite, and large quantities of interstitial chlorite and some specularite were generated. Hydrothermal alteration in the outer zone was marked chiefly by the change from biotite to chlorite, the development locally of clinozoisite, and the generation of interstitial chlorite and a little specularite.

Petrography and Contact Metamorphism of the Vintage Dolomite

Unaltered Dolomite

The fresh, unaltered Vintage dolomite is well exposed along both sides of the north and south quarries. Typically it is a fine-grained rock with a saccharoid texture. In the south quarry, on the east side, the dolomite is gray to bluish gray and possesses a marked foliation caused by the presence of thin layers of oriented mica, a fraction of a millimeter thick. The spacing between mica layers ranges from 2 mm. to 2 cm. Many of these layers are minutely crenulated.

On the west side of the south quarry and on the east wall of the north quarry, the dolomite is nearly pure white and resembles lump sugar. Most specimens are rather massive and lack the good foliation of the gray dolomite, but a few contain thin layers of mica that cause the rock to split readily. The white dolomite is locally spangled with porphyroblasts of brown mica (phlogopite) which lie both parallel to the foliation and across it.

At locality 45, on the west wall of the south quarry, a thick layer of bluish-gray dolomite is interbedded with the pure-white dolomite. This bluish-gray rock resembles exactly that on the east face of the south quarry. The cause of this color difference is not clear, since both rocks have essentially the same grain size and the same mineral composition, but it may be due to the presence of finely divided carbon. Both rocks weather brown or buff, frequently bringing into relief lenses and thin layers of quartz.

In thin section the Vintage dolomite is seen to be typically a fine-grained, foliated rock composed of bands of differing mineral composition that represent original bedding (Pl. 3, fig. 4). The larger bands average 4 to 5 mm. thick and consist predominantly of dolomite with admixed quartz and albite. These thick bands alternate with thinner ones consisting of various mineral assemblages, such as dolomite and phlogopite, phlogopite only, or quartz and feldspar. The layers of phlogopite and those of quartz and feldspar are customarily 0.2 mm. across.

The Vintage dolomite is thoroughly recrystallized and homoblastic. Grains of dolomite, feldspar, and quartz average 0.1 mm. in diameter, but some phlogopite porphyroblasts are nearly 1 mm. across. Modes range widely as indicated below:

	Volume per cent	
Mineral	*Range*	*Average*
Dolomite	45–85	80
Feldspar	2–35	10
Quartz	2–25	5
Phlogopite	3– 7	4
Minor constituents	0– 5	1
		100

The minor constituents are pyrite, magnetite, apatite, zircon, rutile, and tourmaline. Pyrite, the most common, is especially abundant along bedding planes which appear to have served as channels of access.

The mineral dolomite comprises the bulk of each specimen in which it forms elongate xenoblastic crystals. Its average grain size is about 0.1 mm. Index determinations on specimens of unaltered Vintage dolomite from widely scattered points (localities 7, 29, 45) show this mineral to have an omega index of 1.680.

Locally the Vintage dolomite is impure and carries a high percentage of feldspar and quartz. This is particularly true in the vicinity of

locality 27 where the rock contains nearly 45 per cent of these two minerals together with about 10 per cent of other impurities. As indicated previously, the dolomite here is gradational into the Antietam schist to the north. Along the contact between these two rocks, on the west side of the south quarry, and at locality 7 on the east side of the south quarry, the percentage of feldspar and quartz impurities is high and all specimens of dolomite contain at least some of these two minerals. They are somewhat smaller than the dolomite and less elongate, and strain shadows in them indicate that they resisted deformation better than the dolomite. Much of the feldspar shows albite twinning and some shows pericline twinning, but many grains are untwinned. The composition is albite-oligoclase (Ab_{90}). Sporadic grains of microcline and orthoclase were noted in some slides. Due to the similarity in appearance of quartz and untwinned plagioclase, accurate estimates of the relative percentages is impossible.

At localities 11 and 45, in the south quarry, thin laminae and slender lenses of quartz are interbedded with dolomite layers. In these the quartz grains are crudely rectangular with their longer axes, averaging 0.3 mm., lying parallel to the foliation.

Pale-brown phlogopite is an abundant constituent of the Vintage dolomite, and it averages 5 per cent. Most grains show parallel orientation although cross-cutting porphyroblasts are common, some of the latter attaining diameters of 1 mm. Specimens of unaltered dolomite from localities 7 and 45 yielded phlogopite with the optical properties shown in Column 14, Table 1. According to Winchell (1933, fig. 200) this mica belongs to the pholgopite-eastonite series and is rich in magnesia.

Porphyroblasts of biotite, rather than of phlogopite, are found at locality 27 in the north quarry where the dolomite grades into the Antietam schist. These are skeletal and incomplete and have now mostly altered to chlorite. The chlorite is pleochroic in green and pale yellow green, and small, stubby to slender crystals of epidote are scattered through it.

Dolomite of the Contact Aureole

General relations.—Nowhere in the quarry is there a complete, typical section through the contact aureole of the Vintage dolomite that can be studied advantageously. A good section is exposed between localities 5 and 7 in the south quarry, where the aureole is some 65 feet wide, but the alteration here is believed to have been controlled by unusual circumstances. In describing the aureole in the dolomite, therefore, it will be necessary to assemble data from widely separated localities rather than from one continuous section.

Essentially the contact metamorphic aureole consists of two parts: (1) An inner zone, some 20 feet wide, in which thermal metamorphism and later hydrothermal alteration show the greater effects, and (2) an outer zone of variable width where the effects of these two processes are less pronounced. These two zones grade into one another imperceptibly, but for purposes of description it is convenient to consider them as distinct units. Similarly, the outer zone grades into the unaltered Vintage dolomite, and this has been indicated on the geological map (Pl. 1) by an appropriate pattern.

Inner zone.—The typical altered dolomite of the inner zone may be observed at many points along the east wall of the north and south quarries, especially at localities 35, 36, 47, 48, 10, 16, and 19, and at locality 61 on the west wall of the south quarry. It is a poorly foliated to nonfoliated, thoroughly recrystallized, extremely dense, bluish-gray to black limestone, containing fine-grained carbonate and silicates and conspicuous veins of pea-green talc. The rock effervesces freely in cold, dilute hydrochloric acid, indicating that most of the carbonate is calcite. Parallel to the talc veins are thinner layers (probably veins) of calcite which locally take the shape of lenses or irregular masses. Although most bedding and schistosity planes of the original dolomite have disappeared, it seems that these originally served as important channelways along which the talc and the calcite veins were introduced.

At locality 19, near the south end of the south quarry, the fine, black limestone is spotted, rather than banded, with irregular pods of admixed talc and calcite. A thin lenticular slab of altered dolomite that projects into the dike at locality 47 has been changed to a very dense black rock with conchoidal fracture. Thin veins of pale-green, cross-fiber talc fill joints and frac-

tures in the altered dolomite at some places in the inner part of the contact aureole, and thin cross-cutting calcite veins are common. Generally, weathering turns the black limestone light gray or buff.

Microscopically, individual specimens of rock from the inner part of the aureole vary considerably in appearance and percentage mineral composition. However, all show a notable similarity with respect to kinds of minerals, peculiarities of texture, and nature of alteration. Therefore, a general description will best serve to portray the character of these rocks. Due to the wide range in percentage mineral composition, it is meaningless to present modes.

Typically, a thin section of rock from the inner part of the aureole comprises broad bands of carbonate grains with intermingled antigorite (altered from forsterite), phlogopite, and magnetite alternating with thin layers of phlogopite. These bands represent original bedding. Essentially parallel to these are broad, irregular, nearly opaque veins of talc and concentrations of diopside, the latter arranged in radiating rosettes with centers of calcite, tremolite, and antigorite (Pl. 3, figs. 5, 6). Minor constituents include apatite, zircon, rutile (?), epidote, and pyrite, the latter concentrated along stratification planes.

In the inner part of the aureole, most of the dolomite of the original rock has been converted to calcite, although a little dolomite may still remain. This latter possibility could not be established with certainty. The quantity of calcite ranges widely but it is always the dominant mineral of the rock. Its average grain size, in a typical specimen from locality 10, is 0.07 mm., whereas that of the carbonate grains in the unaltered dolomite outside the aureole is 0.1 mm. Adjacent grains of calcite are tightly intergrown and, unlike the original dolomite grains, are nearly equidimensional. Oriented phlogopite flakes, both in layers and as scattered individuals, help the rock to retain some of its original foliation.

Intermingled with the calcite are irregular to ellipsoidal masses of antigorite. This mineral, which is pale brown in plane polarized light, forms aggregates with a fibrolamellar structure. The ellipsoidal outline suggests the subrounded form of olivine crystals, and it is believed that the antigorite has altered from forsterite. Most of the forsterite is gone now, but in one or two slides a small amount remains as unaltered relics. In a specimen from locality 10, pale red-brown fibres fringe an elliptical mass of antigorite, and it is thought that this mineral may be bowlingite, also an alteration of olivine. In some sections, the antigorite appears to be eating into the adjoining calcite and the bounding edges are irregular.

In a number of thin sections minute, colorless scales are associated with antigorite. Their fineness makes positive determination impossible, but they are believed to be talc since all indices of refraction are less than 1.580.

Many of the thin phlogopite layers of the unaltered Vintage dolomite have been preserved in the inner zone of the contact aureole. Although individual plates appear to have grown longer, they retain their original position. Locally the crystals pass into vague, dirty streaks that appear to have been sheared and elongated, and some of these bend around aggregates of calcite. The phlogopite is characteristically clouded with either poikiloblastic inclusions or alteration products. Small flakes of phlogopite are also scattered throughout the calcite-antigorite layers in many specimens. These are well aligned and appear to be relics that have retained their original orientation.

The optical properties of this phlogopite vary slightly in different parts of the inner zone of the aureole (Columns 15 and 16, Table 1). Although perhaps not very significant, they suggest that the phlogopite in the altered dolomite of the inner zone of the contact aureole has a lower ferrous iron-magnesia ratio than the phlogopite in either the unaltered dolomite or the slightly altered dolomite of the outer zone (Winchell, 1933, fig. 200).

One of the most significant contrasts between the altered dolomite of the inner zone and the unaltered dolomite is in the feldspar and quartz content. The dolomite, outside the contact zone, averages 10 per cent feldspar (mostly plagioclase, $Ab_{90}An_{10}$) and 5 per cent quartz as small grains disseminated through the carbonate. With a few unimportant exceptions, these minerals are not present in the inner part of the contact aureole. One section from near locality 5 contains feldspar and quartz

in distinct layers between calcite-antigorite bands, but this occurrence is abnormal and will be described later.

Scattered grains and masses of magnetite are abundant, and in one slide form nearly 8 per cent of the rock. In this specimen the mineral is concentrated in zones parallel to bedding planes, suggesting hydrothermal introduction. It will be recalled that magnetite is merely a minor constituent in the unaltered dolomite outside the contact aureole.

The most striking features of the inner zone of the aureole are the talcose veins which parallel the bedding planes and which clearly have been introduced by solutions. Many of these are elongate, irregular masses; others are more vein-like with approximately parallel walls; still others resemble discontinuous ellipsoidal or lenticular pods (Fig. 3). Their thickness ranges from 0.1 mm. to 1 cm. with an average of 1 mm. It is these that appear as green bands and irregular green lenses in the hand specimen. The outer portion of these veins, and very often the whole vein, is composed of talc, much of which has a fibrous habit. Most fibres are perpendicular to the vein walls, but some are arranged at random forming an intricate network.

The talc was positively identified by George T. Faust of the United States Geological Survey, who, in a personal communication, wrote:

"This identification is based on the similarity of the differential thermal analysis pattern of this mineral with those obtained from several natural talcs and one synthetic talc and on the identity of the powder diffraction X-ray patterns of the same materials. In addition to the talc peak, the thermal analysis pattern shows two small, but well-defined peaks at very low temperatures which are probably due to admixed materials."

In section the talc is brown and nearly opaque (Pl. 3, fig. 5), apparently due to concentrations of fine iron ore particles, so that it appears non-birefracting. However, a specimen from locality 16 was crushed and studied in index liquids and was found to have a birefringence corresponding to that of talc. The aggregate structure makes it impossible to measure accurately the three indices of refraction, but α is about 1.545 and the other indices are between this and 1.580.

In one slide, minute crystals of a colorless to pinkish, birefringent mineral, possibly garnet, are scattered throughout the talc.

Within many irregular talc veins, and particularly within the ellipsoidal or lenticular ones, are small crystals of diopside, some with

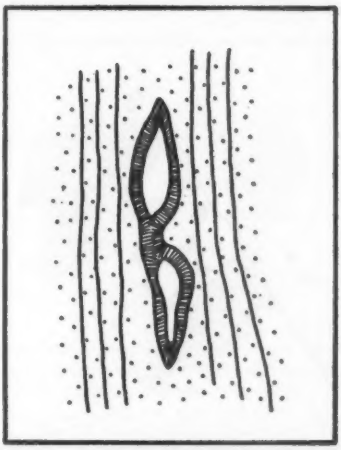

FIGURE 3.—TALC VEIN IN ALTERED VINTAGE DOLOMITE

Stratified Vintage dolomite (stippled) at locality 5 showing vein of talc (shaded) surrounding lenticular cores of calcite (unshaded). Talc vein is about 2.5 inches long.

radial arrangement (Pl. 3, fig. 5; Figs. 4, 5). In some sections, these crystals form solid interiors; in others they merely make linings and the interiors are filled with coarse crystals of calcite or with aggregates of calcite and antigorite. Since some interiors have an outline resembling that of an olivine crystal, it is believed that at least some of these assemblages were built up as reaction rims around early-formed crystals of forsterite. The coarse calcite crystals found in many interiors may represent replacements of forsterite brought about by hydrothermal solutions. Behre, Osborn, and Rainwater (1936, p. 794–796) have described somewhat similar but much larger ellipsoidal and spheroidal masses of diopside with radiating blades replacing limestone in the contact zone at the Calumet iron mine, Colorado.

In several thin sections, sheaf-like bunches of slender, colorless, hair-like fibers are associated with the veins of talc and diopside. Although extremely fine, these fibers appear

to have an extinction angle of about 20° and to show interference colors of low second order. They are believed to be tremolite. The fibers are especially numerous in the rocks from localities 16 and 19 on the east wall of the south quarry. In the latter rock (Pl. 3, fig. 6), dense

FIGURE 4.—CONTACT BETWEEN ALTERED VINTAGE DOLOMITE AND DIABASE APOPHYSIS

Sketch drawn from a thin section of rock from locality 16. Pods of granular diopside (D) are bordered by rims of fibrous talc (T). Small elliptical areas (shaded) are talc in a groundmass of calcite and antigorite (no pattern). Arrow indicating foliation of dolomite is 1 mm. long.

mats of them are surrounded by diopside rings and one obtains the distinct impression that the tremolite antedates the diopside. In another section, the fibers are piercing large crystals of calcite within the diopside rings; in a different part of the same section they are penetrating calcite well away from any diopside.

Outer zone.—The outer zone of the contact aureole may be best studied in the north quarry near localities 25 and 32. In some respects the rock here resembles that in the inner zone of the aureole, in other respects it resembles the unaltered dolomite. Toward the east it grades imperceptibly into unaltered dolomite; toward the west the relations are unknown since much of the rock has been removed by quarrying. Evidence elsewhere, however, indicates that this rock of the outer zone grades into that of the inner zone.

In hand specimen the rock of the outer zone is nonfoliated, sugary-textured, and essentially white, with a high specific gravity and a conchoidal fracture. It effervesces moderately with cold, dilute hydrochloric acid, indicating the presence of some calcite. Irregular areas of pale-green talc are present, and large phlogopite porphyroblasts are concentrated locally.

Microscopically, the rock consists essentially of interlocking grains of carbonate (85 per cent) and irregular areas (15 per cent) of antigorite and talc (?) with scattered flakes of phlogopite and cubes of pyrite. No veins of talc were noted in thin section. The average grain size is somewhat coarser than that of the rock from the inner zone, and many carbonate grains show parallel elongation. Optical properties prove that most of the carbonate is dolomite but some calcite is also present. Antigorite forms both irregular and rectangular masses, the outlines of which suggest that it has altered from forsterite. Talc, the identity of which is not certain, forms minute flakes with diverse orientations, scattered through the antigorite. Quartz and feldspar are entirely lacking. A specimen of slightly altered dolomite from locality 25 yielded phlogopite whose optical properties, and probably also chemical composition, are similar to those of the phlogopite from the unaltered dolomite. The absence of tremolite, diopside, and abundant veins of talc in the outer zone of the aureole suggests that temperatures were probably lower and proves that hydrothermal solutions were less copious than in the inner portion.

Section through localities 5, 6, and 7.—The quarry face between localities 5 and 7 presents a nearly complete section across the contact aureole which, in this place, is some 60 feet wide. Alteration here is believed to have been controlled by shearing and faulting which permitted an unusual opportunity for the migration of hot solutions. Since the character of the alteration along this section is somewhat unique, it is deemed worthy of special consideration.

At locality 5, close to the diabase intrusion, the rock is fine-grained, rudely schistose, speckled with phlogopite porphyroblasts, and distinctly layered with bands from 1 mm. to 1 cm. wide. Some layers are of black carbonate, some of brownish carbonate due to abundant minute flakes of phlogopite, some of talc, and

some of calcite. The rock weathers to alternating gray and green bands. Locally, lenticular pods of pale-green talc lie parallel to the banding of the rock. Some of these contain centers of milky calcite which is younger than the talc (Fig. 5). Calcite incrustations coat many joint faces, and drusy calcite veins, one of which is 4 cm. thick, cut through the rock in random directions. Many of these veins bear pyrite. An irregular quartz vein, 13 cm. thick, is genetically late.

In thin section, the rock at locality 5 shows thin beds of quartz and untwinned plagioclase ($Ab_{90}An_{10}$) between layers of carbonate (chiefly dolomite, some calcite) and antigorite and thin layers of phlogopite. Many of these beds are sheared and pulled out into lenticular bodies or sliver-like augen surrounded by carbonate. Apparently the carbonate flowed under stress, permitting the quartz-plagioclase layers to shear. Bands of phlogopite bend gracefully around some of these quartz-plagioclase augen, and in one place, large flakes of phlogopite were wrapped about lenticular eyes of carbonate. In spite of its position near the contact, this rock contains neither tremolite nor diopside.

At locality 6 a zone of shattering, probably caused by faulting, trends N 20° E and is essentially vertical (Pl. 2, fig. 3). The rock here is broken and highly replaced by masses of talc. Many veins of calcite and fibrous talc, 3 mm.–4 cm. thick, cut through the more massive talc in random directions. Modified tetrahedrons of black sphalerite, one about 12 mm. in diameter, were found embedded in talc and calcite veins. The sphalerite gave tests for both zinc and sulphur, and its index of refraction was determined by Jewell Glass of the United States Geological Survey to be 2.370 ±0.005. The mineralization at locality 6 is clearly the result of hydrothermal solutions moving along a zone of shattering.

Microscopic study shows that the rock from locality 6 contains augen and sliced lenses of plagioclase similar to those at locality 5. Constituent minerals are calcite, dolomite (?), antigorite, phlogopite, tremolite, diopside, and talc, those characteristic of the inner zone of the aureole. Although more than 40 feet southeast of the dike contact, this rock has been rather intensely altered.

Fifteen feet south of locality 6, the altered Vintage dolomite is banded with dark-gray, light-gray, brown, and green layers. These are all parallel to the original beds which are here thrown into rather close folds plunging S 80° W at an angle of 25°. Here, too, calcite veins abound along bedding planes and fractures, and

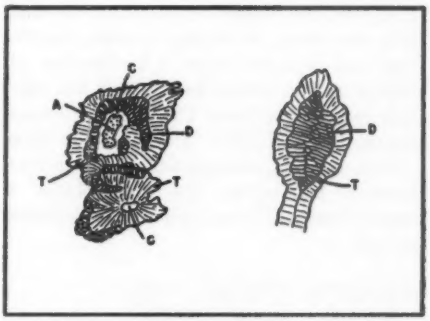

FIGURE 5.—HYDROTHERMAL MINERAL ASSEMBLAGES IN ALTERED VINTAGE DOLOMITE

Sketches drawn from a thin section of dolomite from locality 19. Figure on left shows rims of diopside (D) and talc (T) surrounding earlier cores, probably originally forsterite, now altered chiefly to antigorite (A) and partly replaced by calcite (C). Figure on right shows a center composed of diopside granules (D) inclosed by a rim of talc (T). Each assemblage is about 1.8 mm. long.

slickenside calcite has formed along joints. Several small quartz veins were observed cutting calcite veins, demonstrating that the quartz is younger.

Southward, toward locality 7, the rock of the outer zone grades imperceptibly into unaltered dolomite, but outcrops do not permit detailed study.

Contact Metamorphism of the Dolomite

The unaltered Vintage dolomite at Safe Harbor is a white, grayish, or bluish rock with a fine-grained saccharoid texture and marked foliation. Its chief constituents are dolomite, phlogopite, quartz, and feldspar. It was into this rock that the diabase dike was intruded, transecting the bedding and foliation planes at a high angle. The resulting contact metamorphism took place in two stages: (1) A thermal stage during which the dolomite throughout the full width of the aureole was

altered ("contact" pyrometamorphism of the economic geologist), and (2) a hydrothermal stage during which hot solutions from the diabase reworked the rocks of the whole thermal zone, but most intensely those near the dike contact ("contact" pyrometasomatism of the economic geologist).

In the thermal stage, the advancing wave of heat from the diabase recrystallized the dolomite to a dense, bluish-gray to black, massive rock and destroyed most of its foliation, although in thin section remnants of this foliation can still be seen. The earliest chemical change was a reaction throughout the whole aureole between dolomite and quartz to form forsterite and calcite, with the evolution of carbon dioxide. This reaction may be illustrated as follows:

$$2\,CaMg(CO_3)_2 + SiO_2 = 2\,CaCO_3 + Mg_2SiO_4 + 2CO_2$$
(dolomite) (quartz) (calcite) (forsterite)

Although the forsterite is now largely altered, its former presence is revealed both by small patchy remnants of itself and by the outline of its alteration products. With the production of forsterite, most of the carbonate was converted to calcite and the black contact rocks now effervesce freely in cold hydrochloric acid. This process, known as dedolomitization, is common in contact metamorphism and results from the fact that the magnesium carbonate by itself is more easily dissociated than the calcic.

At this stage of alteration forsterite, the orthosilicate, rather than tremolite or diopside, formed since the amount of silica in the Vintage dolomite was limited. Although most of the silica was undoubtedly furnished by quartz, plagioclase may have contributed a small amount. Except near localities 5 and 6 where conditions were somewhat abnormal, all quartz and nearly all plagioclase have disappeared from the zone of contact alteration.

In the altered Vintage dolomite, traces of original bedding are still marked by thin layers of parallel phlogopite flakes. Many of these appear dirty and streaked out and seem to have recrystallized and grown longer. Their indices of refraction suggest that their ferrous iron/magnesia ratio has decreased in the process, but whether this change was effected during the thermal stage or during the later hydrothermal stage is not known.

It is reasonably certain that the development of lime-silicates in the Vintage dolomite took place in the following sequence: forsterite, tremolite, diopside. Forsterite, the orthosilicate, is disseminated evenly throughout the altered rock where it formed during the thermal stage. Tremolite and diopside, on the other hand, are confined almost exclusively along bedding and foliation planes, and it is fairly clear that they are genetically related to hot aqueous solutions that circulated in the hydrothermal stage. Analogy with the metamorphism of the schist suggests that the temperature peak had passed when solutions appeared and that both tremolite and diopside formed at lower temperatures than forsterite. Tremolite and diopside are metasilicates, requiring more silica than forsterite, and the incoming solutions seem to have brought in the necessary silica for these two minerals and the necessary water for tremolite. The reactions cannot be expressed adequately by balanced equations for they probably took place by an exchange of ions between the reacting substances.

The appearance of large amounts of talc marked the climax of hydrothermal alteration. Temperature had now abated considerably, and solutions were most abundant. Bedding and foliation planes, fault planes, and shear zones clearly served as channels of access because the talc is confined principally to these zones. It seems mandatory in this case that the production of most of the talc was occasioned by the introduction of both magnesia and silica into the contact zone. However, the magnesia may have come from the dolomite at depth and the silica may have been extracted from the adjacent schist at depth. Iron was brought in at this time also, and magnetite became a distinctive constituent of the rock, especially along bedding planes.

Talc and magnetite are confined mainly to the inner zone of alteration and it has already been noted that tremolite and diopside are also confined here. Thus, up to this point, the hydrothermal effects of the dike intrusion were felt most strongly in the inner zone of the aureole.

As the temperature dropped and the solu-

tions penetrated farther into the country rock, forsterite was altered to antigorite throughout the entire width of the aureole, a change involving the addition of both silica and water. This change may be expressed as follows:

$$3\,Mg_2SiO_4 + SiO_2 + 4\,H_2O \longrightarrow 2\,Mg_3Si_2O_5(OH)_4$$
(*forsterite*) (*antigorite*)

At the same time these siliceous solutions produced an endomorphic effect on the diabase dike near its contact, altering some phenocrysts of olivine to antigorite and hydrating the tachylitic borders to chlorophaeite.

At locality 6, the metamorphism was influenced greatly by a wide zone of shattering that served as a convenient channel along which solutions could move from the dike. Thus the alteration here was more intense than might be expected in view of the distance from the contact. Forsterite was first formed, followed by tremolite, diopside, talc, and sphalerite, and by the conversion of forsterite to antigorite. Veins of quartz and calcite concluded the hydrothermal stage here.

The succession of metamorphic events in the Vintage dolomite followed in general the same pattern as that in the Antietam schist; i.e., thermal metamorphism followed by hydrothermal metamorphism. In the dolomite, however, the thermal alteration extended across the complete width of the contact zone diminishing in intensity away from the intrusion. The rock was recrystallized and forsterite was the first mineral to develop, with resulting dedolomitization of the rock, and the disappearance of quartz and feldspar especially near the contact. With the passing of the temperature peak, hydrothermal solutions appeared, bringing in silica to form tremolite and diopside, especially near the contact. This mineralization was followed by the introduction of large amounts of magnesia and silica to form talc, iron to form magnetite, and, at locality 6, small amounts of zinc and sulphur to produce sphalerite. Up to this point hydrothermal changes were confined mainly to the inner part of the aureole, except at locality 6 where the sheared rock facilitated the lateral movement of solutions. With declining temperature, forsterite in both the inner and outer zones became unstable and under the influence of hot silica-bearing water was changed to antigorite. The thermal effect of the diabase was felt through a wider zone in the dolomite than it was in the schist. In both rock types, however, hydrothermal solutions penetrated and recrystallized the whole contact zone, but more intensely that part lying close to the intrusion.

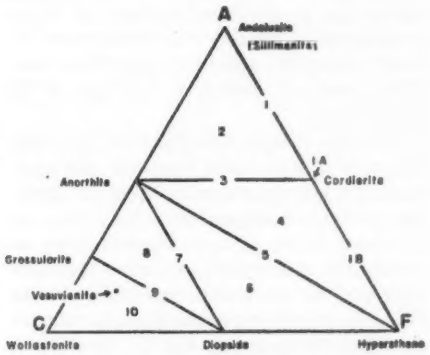

FIGURE 6.—PYROXENE HORNFELS FACIES
This ACF diagram applies to rocks with excess silica. Modified from V. M. Goldschmidt (1911).

EQUILIBRIUM RELATIONS DURING METAMORPHISM

If we judge from the mineral assemblage developed by thermal alteration, then the Antietam schist, prior to hydrothermal metamorphism, belonged to the pyroxene hornfels facies, a diagrammatic representation of which is shown in Figure 6. This ACF diagram is modified from the one prepared by Goldschmidt (1911) for the hornfelses. With excess silica, several classes of quartz-bearing hornfels are possible in this facies, depending upon the original bulk composition of the rock. Class 3, for example, lies on the anorthite-cordierite join and represents a mineral assemblage of cordierite and plagioclase with quartz, orthoclase, and biotite as possible but not critical members of the association. The thermally altered Antietam schist belonged to this class because during the thermal stage it was recrystallized to an aggregate of cordierite, plagioclase, orthoclase, and quartz in a zone immediately adjacent to the intrusion. Both muscovite and biotite are notably absent from this

rock and it seems likely that equilibrium conditions were attained here. However, in the outer part of the thermal zone, i.e., 4 or 5 feet from the contact, it is evident that the rock was less recrystallized and that mineral transformations were much less complete, nascent cordierite showing poorly defined outlines and biotite porphyroblasts remaining as relicts (now altered to chlorite). Apparently equilibrium conditions were not reached in this part of the thermal zone.

Since the Vintage dolomite must have been thermally altered under essentially the same temperature-pressure conditions as the Antietam schist, it is isofacial with the latter, i.e., it belongs to the pyroxene hornfels facies, although this conclusion is not necessarily substantiated by the mineral assemblage of the dolomite. Late hydrothermal alteration has confused the picture greatly, but it is clear that in the thermal stage, forsterite and calcite were first produced at the expense of dolomite and quartz. Bowen (1940, p. 240) has pointed out that it is normal for forsterite rather than the metasilicate enstatite to form as the first phase in the thermal metamorphism of siliceous dolomite, but that the reaction is not an equilibrium reaction because the phase assemblages involved include forsterite, a phase which individually cannot exist in equilibrium with dolomite and quartz. In the present case, all available silica appears to have been used up in the reaction and, as pointed out previously, it is quite possible that a little dolomite may still remain in the rock.

Bowen (1940, p. 260) has arranged the mineral phases produced in the thermal metamorphism of siliceous dolomite in the order of their production with rising temperature: These are:

1. Tremolite
2. Forsterite
3. Diopside
4. Periclase
5. Wollastonite
6. Monticellite
7. Akermanite
8. Spurrite
9. Merwinite
10. Larnite

Forsterite forms at a higher temperature than tremolite, and diopside at a higher temperature than forsterite, so that at first sight it might be thought that this arrangement could be used as a general indicator of rising temperature. Bowen emphasizes, however, that this sequence applies strictly to the thermal metamorphism of carbonate rocks and that if circulating solutions are present they could introduce a new phase or even induce its formation at temperatures far below that at which it could be produced in thermal metamorphism.

The metamorphic sequence in the dolomite at Safe Harbor is forsterite, tremolite, and diopside, an arrangement quite different from that given above. Forsterite seems to have formed first under thermal conditions when the rocks were hot, dry, and silica-deficient. Later, with lower temperature and the influx of silica-bearing solutions, forsterite became unstable and tremolite and diopside developed.

Bowen (1940, p. 255–264) has also emphasized the importance of pressure on the stability of mineral assemblages produced in the alteration of carbonate rocks such as the Vintage dolomite. All the steps exhibited by the above list of minerals are univariant reactions that result in the evolution of CO_2, and the temperature at which they take place depends upon the applied pressure of CO_2. Bowen has drawn pressure-temperature curves which show that low pressure acts in the same direction as high temperature, and that at a given temperature a lower pressure will favor the development of higher members of the series. Conversely, under the same temperature conditions, a higher pressure will favor development of lower members. Now the formation of lime-silicates involves the evolution of large amounts of CO_2 and if the rock is relatively impermeable this will build up a high pressure, thus favoring the development of the lower members of the series. Later, if this CO_2 were, by chance, to pass off as a gas or to be dissolved by hot, circulating solutions and thereby removed from the system, the resulting reduction of pressure would permit higher members of the series to form. This may explain the relatively late formation of diopside in the Vintage dolomite in the hydrothermal stage, when temperature seems to have been on the wane. Furthermore, it may also satisfactorily account for the relatively high-temperature alteration at locality 6, a comparatively long distance from the dike contact. The CO_2, either as a gas or in a solution, could have moved freely along the fracture zone passing through this point, thereby

lowering the pressure and permitting diopside to form.

Since both the schist and the dolomite, before hydrothermal alteration, belonged to the pyroxene hornfels facies, they must have been thermally metamorphosed under high temperature and probably moderate pressure, but it is impossible to estimate accurately what the maximum temperature may have been. Bowen (1940, p. 263) has pointed out that in metamorphism the highest members of the lime-silicate series are associated only with basic rocks and especially with hypabyssal intrusions. Larsen (1929, p. 94) has shown that the temperatures of most gabbroic magmas probably range between 800° and 900° C., and Turner (1948, p. 70) has estimated that 700° to 750° C. is not too low a value to place upon the temperature of transition between the pyroxene hornfels and the next lower (amphibolite) facies. Perhaps then, the temperature of thermal metamorphism at Safe Harbor lay somewhere around 800° C. Whatever the maximum temperature may have been, the distribution of the thermal aureole suggests that a high temperature was reached rapidly, adjacent to the contact, maintained for a brief time, and then declined rather quickly at about the time when hydrothermal solutions appeared.

Comparison with Other Areas

Purpose

Comparison of the metamorphism caused by the Safe Harbor dike with that caused by similar igneous bodies in other areas will enable us to draw certain conclusions regarding the physico-chemical conditions which prevailed adjacent to these basic hypabyssal intrusions. The Palisades sill of New Jersey aud the great Whin sill of northern England are selected for comparison.

Palisades Sill

The Palisades sill is a great sheet of intrusive diabase, hundreds of feet thick, that outcrops for about 100 miles across New York and New Jersey. Andreae and Osann (1892) and Irving (1899) investigated the contact-metamorphic effects of this diabase on Triassic shales and arkoses in New Jersey, but the most detailed examination was made by Lewis (1907), who pointed out that the sediments here were "baked" by the magma during the prolonged stages of its cooling under deep cover. He observed:

> "This is particularly true of the shale, which constitutes the most abundant constituent of the Newark formation in New Jersey. From the contacts, outward, through a thickness of several hundred feet, the shale has been everywhere changed into a hard flinty gray to brown and black hornfels, having the hardness of slate but lacking its splitting qualities, and the original lamination is preserved only in the banding of the colors." (p. 138)

Lewis found that the mineral constitution of the dense aphanitic hornfels varies widely and includes many different combinations of feldspar (both orthoclase and plagioclase), biotite, augite, hornblende, tremolite, garnet, spinel, magnetite, quartz, muscovite, cordierite, scapolite, vesuvianite, sillimanite, andalusite, chlorite, calcite, analacite, titanite, tourmaline, zircon, apatite, and possibly leucite. He divided the hornfelsic rocks into a number of types which he named from the most prominent or characteristic mineral constituent, such as biotite-hornfels, chlorite-hornfels, and lime-silicate hornfels.

The Palisades diabase sill also intruded arkosic rocks (feldspathic sandstones), but these were less susceptible to alteration. The metamorphic zone is much narrower than in the shale and only a few feet from the contact the rock shows no visible effect. The altered arkose now contains orthoclase and plagioclase in varying proportions, together with varying quantities of quartz, augite, biotite, epidote, cordierite, chlorite, calcite, tourmaline, and apatite. Lewis recognized several groups of arkose which he named from the prevailing or characteristic constituents, such as augite-arkose, epidote-chlorite-arkose, and cordierite-arkose.

Judging from Lewis' description, the cordierite-arkose resembles the altered Antietam schist at Safe Harbor. He described this rock as

> "a fine grained orthoclase-arkose, considerably kaolinized and thickly set with cordierite in rectangular sections and pseudo-hexagonal trillings. The mineral

is largely altered to confused aggregates of muscovite (pinite) and granules of magnetite, but some pale yellow to colorless crystals and numerous remnants of unaltered cordierite still remain." (p. 147).

It is difficult to compare the contact metamorphism at Safe Harbor with that adjacent to the Palisades sill, since the original rocks in the two areas were different. In New Jersey, they were mainly unmetamorphosed shales and arkoses. Although originally both of these were mineralogically different from the Antietam schist, their chemical compositions may have been rather similar. Judging from Lewis' microscopic description of the varieties of unaltered arkose, the essential minerals were orthoclase and plagioclase, with or without quartz, and there were probably also minor constituents like mica, chlorite, and magnetite. If combined in a different way, the oxides in these minerals could give rise to a rock like the Antietam schist in which potash, alumina, and silica are contained in muscovite and biotite rather than in orthoclase.

Lewis believed that the arkoses were subjected to high temperatures during alteration and pointed out that the original feldspar was very stable under these conditions. He makes no mention of muscovite in any of the altered arkoses, so if any was there originally, it must have become unstable. The characteristic association of such minerals as cordierite, pyroxene, plagioclase, quartz, orthoclase, and biotite, and the absence of muscovite in the altered arkoses adjacent to the Palisades sill suggest that the rocks belong to the pyroxene hornfels facies and that during metamorphism the temperature-pressure conditions in some parts of the contact zone were similar to those at Safe Harbor.

Whin Sill

The Whin sill is a huge intrusion of late Carboniferous or early Permian quartz dolerite trending for over 100 miles through Northumberland and Cumberland in northern England. Its thickness ranges from only a few to 240 feet with an average of nearly 100 feet. Throughout the greater part of its extent the sill lies within the heart of the Carboniferous Limestone Series which consists mainly of limestones, shales, and sandstones. Locally it transects the bedding, thus coming into direct contact with all these rock types.

The contact aureole, which at one point is over 80 feet wide, was described in detail many years ago by Hutchings (1898). He found that the action of the sill on nonargillaceous limestone with quartz impurities was limited to recrystallization. He stated: "Calcite and quartz recrystallized side by side, and it is rare to see in these particular rocks any formation of wollastonite, or of any calcareous hornfels-like products, such as are more frequently encountered round granite-contacts" (p. 73).

Argillaceous limestones, however, were not only recrystallized to a distance of 60 feet from the contact but show a great variety of new minerals, including garnet, augite, idocrase, wollastonite, epidote, hornblende, feldspar, chlorite, and sphene. Garnet is persistent, whereas the other minerals are present in some cases and absent in others. As in the non-argillaceous limestone, wollastonite is uncommon.

Where the contact action was most intense the purest shales were indurated, bleached, and completely recrystallized, with the formation of a new mass of white mica throughout, well-developed flakes lying crisscross in all directions. Chlorite is intermittently interwoven with the mica, forming spherulitic groups. Newly recrystallized quartz is common and other minerals include feldspar, biotite, andalusite, and anthophyllite. Sandy shales were altered in a somewhat similar manner.

The calcareous shales, according to Hutchings were the most intensely altered of any. They are now brown, compact, hornfelsic rocks with a jaspery appearance and fracture. In thin section they have a fine-grained groundmass of quartz and feldspar in which are embedded crystals of garnet, idocrase, wollastonite, sphene, epidote, augite, hornblende, calcite, and incipient cordierite.

Pure sandstones were simply recrystallized, but argillaceous sandstones now contain mixtures of muscovite, chlorite, and biotite.

Hutchings also determined, from a study of chemical analyses, that during metamorphism soda was added to at least some of the sediments by the magma.

Hutchings' mineral descriptions are too gen-

eral to permit positive assignment of the rocks to a particular facies. However, in the purest shales the association of muscovite (with chlorite), feldspar, biotite, andalusite, and anthophyllite certainly suggests that these rocks belong to the amphibolite facies rather than to the pyroxene hornfels facies. If this is true, then the temperature of metamorphism was lower here than adjacent to either the Palisades sill or the Safe Harbor dike. Hydrostatic pressure adjacent to the Whin sill was certainly no more than moderate, so that recrystallization of quartz and calcite, side by side, in the siliceous limestones is further evidence that the temperature was not high.

Conclusions

Both Lewis and Hutchings described their altered rocks in a general way without stating specifically what minerals are associated in individual specimens. Neither writer compared the mineral assemblage in one part of the aureole with that in another. Lewis merely named his various types of hornfels from the most prominent or characteristic mineral present. In speaking of these types he said:

"It should be clearly understood, however, that these are not sharply defined types, but present various degrees of gradation from one to another. Furthermore, they do not form zones or belts in any consecutive order or other systematic relation to the intrusive rock, but alternate irregularly throughout all parts of the zone of metamorphism. It is evident, therefore, that the several types of hornfels are not the result of varying degrees of metamorphism, but are dependent only on original differences in the composition of the shales themselves." (Lewis, 1907, p. 139-140).

Hutchings seems to have felt that the mineral assemblages adjacent to the Whin sill were also the sole result of original composition. Neither writer considered that distance from the contact might be an important factor, and thus they overlooked the significance of temperature.

In spite of these difficulties it is possible to make a few general comparisons regarding metamorphism in the three areas:

1. The rocks cut by the Palisades sill and Whin sill were unmetamorphosed sediments, whereas those intruded by the Safe Harbor dike were regionally metamorphosed schist and dolomite of medium grade. Through thermal alteration, rocks of similar original bulk composition in the three areas converged toward a common metamorphic type.

2. So far as can be ascertained from Lewis' account, the thermally altered rocks around the Palisades sill, like those at Safe Harbor before hydrothermal changes, belong to the pyroxene hornfels facies and were metamorphosed under high temperature and probably moderate pressure. The evidence suggests that the sediments around the Whin sill now belong to the amphibolite facies and that they were altered at moderate temperature and pressure.

3. Lewis did not mention the introduction of any material into the country rock, but Hutchings found that soda was added to at least some of the sediments. In the contact zone at Safe Harbor, water, ferrous iron, and magnesia, and possibly alumina and silica were added to the schist, and water, magnesia, silica, iron, zinc, and sulphur were brought into the dolomite. The magnesia and silica, however, were probably not of magmatic origin.

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ORE MINERALIZATION IN THE HOMESTAKE GOLD MINE, LEAD, SOUTH DAKOTA

By JAMES A. NOBLE

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ABSTRACT

The Homestake ore bodies are spindle-shaped or pipelike replacements of the pre-Cambrian* Homestake formation where second-stage folds, or cross folds, intersect earlier isoclinal folds. Localization of ore in the cross folds is believed to be due dominantly to the permeability produced by dilatation of the previously well oriented schists.

Field relations and microscopic study enable subdivision of Homestake mineralization into four perhaps interrelated stages. The principal minerals in the first stage are quartz, chlorite, and arsenopyrite; in the second stage, quartz, ankerite, and pyrrhotite. In the third stage only pyrrhotite was deposited. In the fourth stage, following the intrusion of the Tertiary rhyolite dikes, were deposited mostly pyrite and calcite, and minor amounts of a great variety of other minerals. The four stages are separated by structural breaks, at first measured by very minor fracturing but between the third and fourth stages by widespread fracturing, faulting, and dike intrusion. Gold can be interpreted to have been deposited during the fourth stage but superimposed on the products of mineralization of the three earlier stages.

Several unanswered questions remain. The age of the gold mineralization cannot be established. It may be predominantly pre-Cambrian, wholly Tertiary, or partly of each age. It is not known why practically all the gold is confined to the Homestake formation although some of the adjacent formations differ very little from the Homestake in physical and even in chemical character. Finally, an answer to the question of why the mineralization occurred in a sequence of stages is needed not only for this deposit but for most

* This spelling is retained in this paper in the interest of consistency with earlier papers in this series.

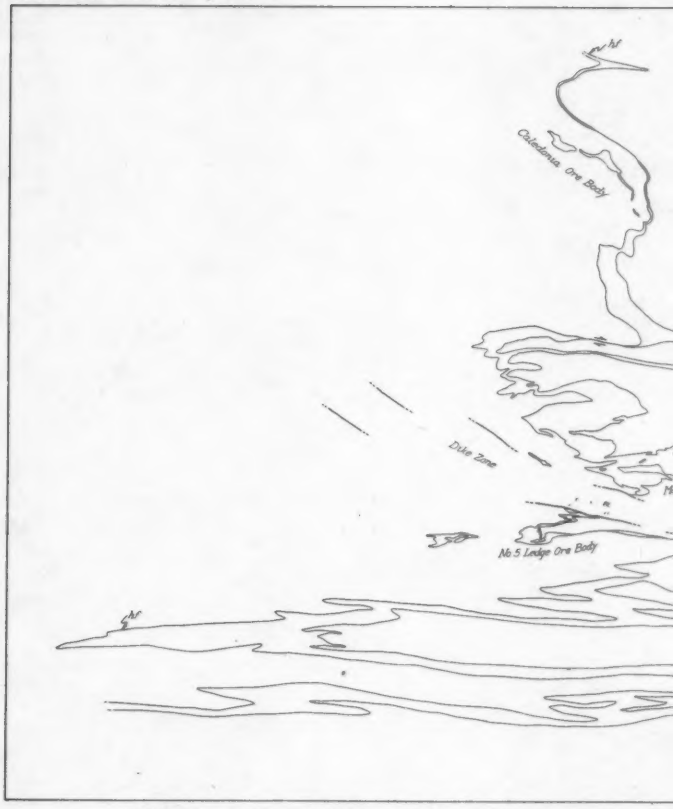
ore deposits. The question is raised whether the widely accepted depth-zone classification of ore deposits should not be replaced by a classification based on composition of the ore-forming fluids.

Introduction

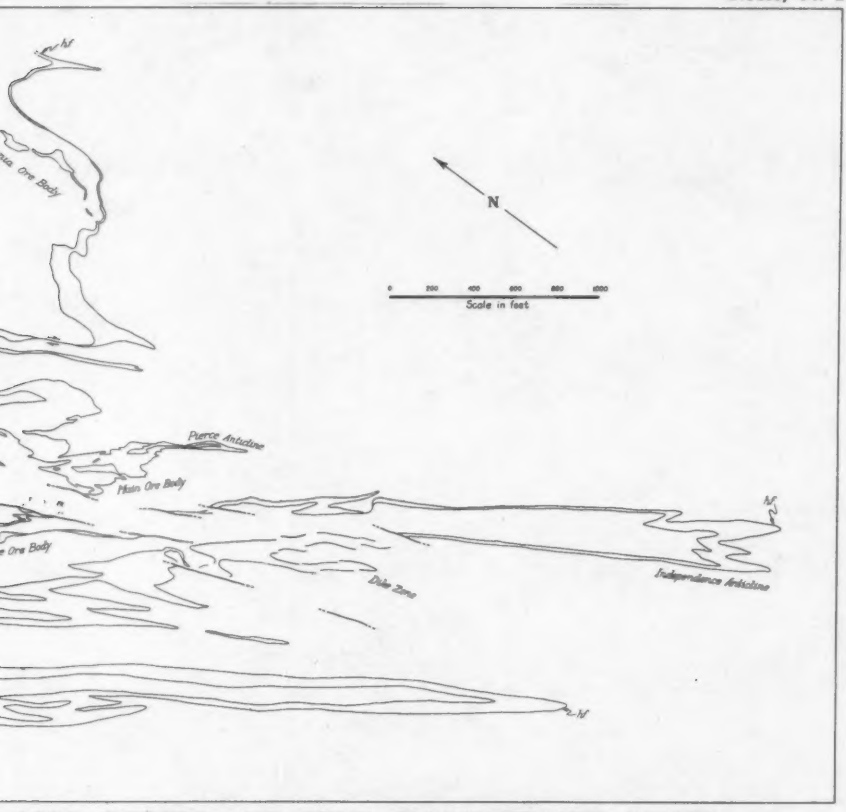
The discovery of the Homestake gold deposit in the spring of 1876 was at the time merely an incident in the exploration and prospecting of the Black Hills, and its full importance in establishing a great industry was not immediately realized. The first record of gold in the Black Hills had previously been made in the southern Hills in the summer of 1874, by prospectors attached to Custer's reconnaissance expedition (Ludlow, 1875). There were rumors of gold in the Black Hills prior to Custer's expedition, however, and small amounts of gold may have been found earlier by fur traders and lone prospectors. The gold found in the southern Hills by the prospectors of the Custer expedition was placer gold freed by the erosion of small veins of gold-bearing quartz in the pre-Cambrian schists. Neither the placers nor the gold-bearing quartz veins were rich enough to work profitably at that time. In fact, neither the placers nor the veins have produced any large amount of gold or other minerals since then; but the mere fact of the presence of gold was sufficient to bring prospectors swarming over the entire Black Hills. Within 2 years there were discovered the really rich placer deposits of Deadwood Gulch and of Nigger Hill, the Homestake deposit, which was at first explored as a series of separate ore bodies, the "cement ores" in the Cambrian conglomerates, and the replacement ores in the Cambrian strata in the Bald Mountain district. The working of the rich placers in Deadwood Gulch apparently led directly to the discovery of the Homestake deposit, but the connection was accidental. It is obvious from the distribution that the placer gold was derived not from the outcrop of the Homestake deposit but from the flat-lying deposits of coarse gold in the Cambrian conglomerates.

Very little was written about the geology of the Homestake mine during the first 25 years of the life of the mine, and almost no geologic record of those years is preserved. In the first few years after the discovery of the deposit, several mining engineers made reports to the officers of the group of companies that eventually were consolidated into the Homestake Mining Company, but all but one of these reports seem to have been destroyed in the San Francisco fire. There is preserved one report by Janin (1879), written only a couple of years after the discovery of the ore bodies. The report and accompanying maps are of historical interest but give only a brief description of the mineralogy, a statement that the porphyry dikes cut the ore, and a conclusion that the presence of gold in the Cambrian conglomerates is proof of pre-Cambrian age of the Homestake ores. Devereux (1882) briefly described the deposits in the Cambrian conglomerate, which even at that early date were nearly worked out. The gold was in his opinion unquestionably placer and the deposits were "fossil placers"; the source of the gold he concluded was the Homestake deposit, which therefore must have existed in pre-Cambrian time. Several writers repeated Devereux's statements and conclusions within the next few years. On the other hand, Hewett (1903) stated that, because all the Tertiary porphyries of the northern Black Hills were gold-bearing to some degree, the ore deposits, including the Homestake deposit, were probably of Tertiary age.

Throughout this time access to the Homestake mine by geologists seems to have been difficult. However, when a study of the mineral deposits of the northern Black Hills was undertaken by the U. S. Geological Survey (Irving, Emmons, and Jaggar, 1904), permission was granted to study the Homestake deposit. Emmons studied the Homestake mine, and Irving made a petrographic study of a few ore specimens. Although permission to make the examination was withdrawn before the work was completed, the descriptions of the structure and mineralization were of some value. The fact that pyrrhotite was identified only doubtfully in a single specimen will be discussed. It was concluded that there had been both pre-Cambrian and Tertiary mineralization; the evidence for the pre-Cambrian mineralization was the presence of gold in the Cambrian conglomerates. Irving decided that the conglomerates contained both placer gold and introduced gold of Tertiary age.



MAP OF PART OF THE 2300 L
Deformation of the Homestake formation (hf



2300 LEVEL OF THE HOMESTAKE MINE

rmation (hf) is a result of isoclinal folding and cross folding

The next paper of geologic interest (Sharwood, 1911) gave analyses of a great many rocks, minerals, and mine waters from the Homestake mine. Although the geologic descriptions were meager, the data were exact and of utmost importance and are still of great value. Several metamorphic and hydrothermal minerals were first identified in this paper. Sharwood did not discuss the age of the Homestake deposit, but some of the facts he presented can be used to support an argument for Tertiary age: the pyrite in the rhyolite dikes carried some gold; a vug in a calcite vein contained fluorite, coarse gold, and a bismuth telluride; and, in the lower levels of the mine, pyrrhotite increased at the expense of pyrite, and the silver-gold ratio decreased.

Paige (1913) was given permission to study and describe the structural relations of the Homestake mine. His conclusion that the ore body was a replacement of a folded sedimentary formation still ranks as the most important single step in the development of geologic theory of the Homestake deposit. He was mistaken, however, in believing that the ore was localized by a fault.

The Homestake Mining Company established a geological department in 1920, with Hosted and Wright as the first geologists. Shortly after this Paige joined Hosted and Wright in a comprehensive study of the Lead area (Hosted and Wright, 1922; 1923; Paige, 1923; 1924; Darton and Paige, 1925; Wright, 1926). Hosted and Wright quickly became convinced that the mineralization which included the gold was younger than the dikes and was confined to a zone a few hundred feet wide with the dikes in the center. They concluded that the gold-bearing portion of the mineralization was introduced by the rhyolite intrusion. Paige thought that some gold was younger than the dikes but that most was older. The older part he thought was pre-Cambrian because mineralization began before the regional metamorphism had ended. Moore (1925), after a brief visit, supported the arguments of Hosted and Wright. He added that some of the gold-bearing portion of the ore was post-rhyolite because it cut the dikes, but this statement cannot be supported by any facts now known.

Connolly (1927) published results of several years of study of the gold deposits of the northern Black Hills, which included a microscopic study of the Homestake ores. He identified some new minerals and distinguished regional and hydrothermal minerals. Connolly concluded that the main mineralization was pre-Cambrian but a part, perhaps including some gold, was Tertiary. McLaughlin (1928), reviewing Connolly's paper, concurred.

Several studies have recently been made of separate features of the Homestake deposit. Some have been embodied in undergraduate or doctorate theses, and others have been published. Gibson (1928) studied a large collection of Homestake specimens with the reflecting microscope and noted several new minerals and new relations. Gustafson (1930; 1933) differentiated and described the regional and hydrothermal minerals in the Homestake formation. Kutz (1930) studied a suite of ore specimens with the reflecting microscope and described the mineral relations. He decided there was only one period of mineralization—Tertiary. Yates (1931) discussed the structure of the Homestake mine and the localization of the ores. McLaughlin (1931; 1933) restated and summarized the results of the studies of Gibson, Gustafson, and Yates and gave the results of his own studies of the field relations. Smith (1935) made a special study of mineralization products in and alongside the rhyolite dikes but found no conclusive evidence bearing on the age of the gold. Wright (1937) amplified the arguments supporting the Tertiary age of the gold. He concluded that some quartz, arsenopyrite, and gold were pre-Cambrian, but more quartz, arsenopyrite, pyrrhotite, pyrite, and most of the gold were Tertiary.

The writer is indebted to individuals almost too numerous to mention for critical discussion of the geology of the Homestake ore bodies, but the principal indebtedness is to the Homestake Mining Company for opportunity to carry on all the phases of the work and for permission to publish the results. D. H. McLaughlin, formerly consulting geologist and now president of the Homestake Mining Company, has had a part in the work represented by this paper from the very concept of the ideas to the reading of the finished manuscript, and he has participated in the final adjustment of most

of the theories without, however, being responsible for them. The drawings were made by Joan T. Rounds. A preliminary draft of the manuscript was read critically by G. M. Schwartz, J. W. Gruner, R. G. Wayland, and N. Herz, and the final manuscript was read critically by J. O. Harder, A. L. Slaughter, and G. C. Mathisrud. Few visitors to the Homestake mine have missed a discussion of the theory of the structural control of the ore deposits, and, without doubt, some of the theory in its present form has resulted from these discussions.

The Homestake mine is located in the northern part of the Black Hills, an elongated dome in western South Dakota and adjacent parts of Wyoming and Montana. Pre-Cambrian schists are surrounded by outward-dipping Paleozoic and Mesozoic rocks. Some pre-Cambrian pegmatitic granite is exposed in the southern part of the Hills, but there is very little in the northern part and none in the immediate vicinity of the Homestake mine. Tertiary intrusive rocks, on the other hand, are fairly abundant in the northern Hills but are not exposed in the southern part. Gold has been mined from widely scattered areas throughout the Black Hills uplift, principally from the pre-Cambrian rocks but also from parts of the immediately adjacent Paleozoic rocks and from Tertiary and Recent gravels. Production from the Homestake mine, both past and current, far exceeds all the other sources.

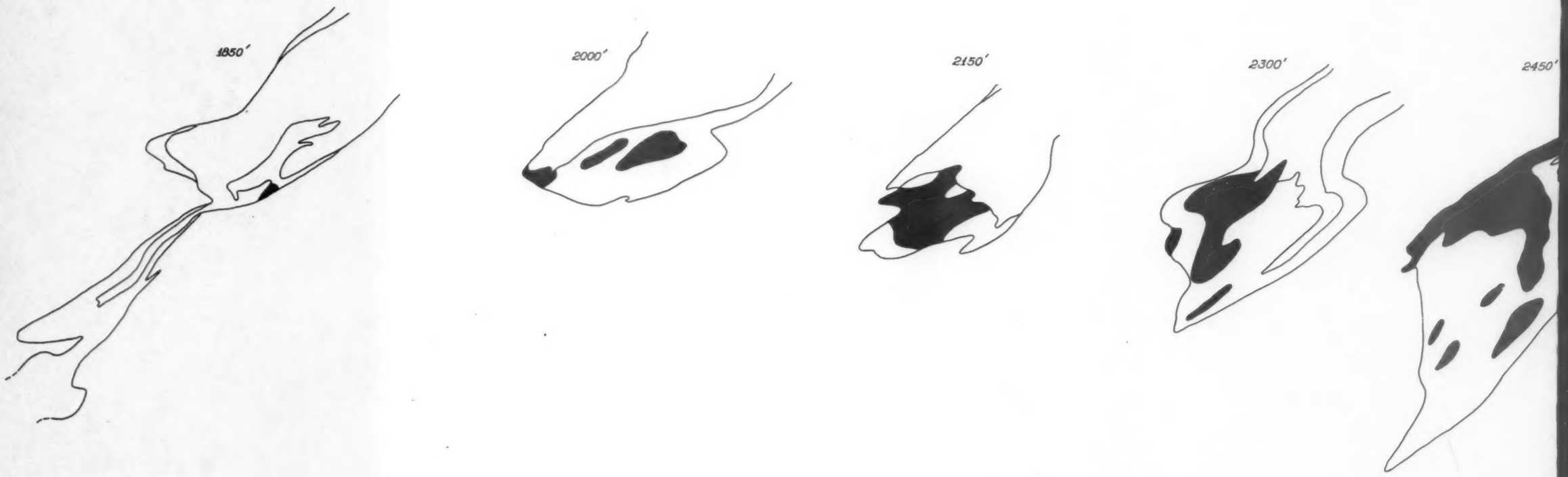
The pre-Cambrian schists in the vicinity of the Homestake mine are metamorphosed sedimentary rocks, dominantly iron-magnesium carbonates in the lower part of the sequence and dominantly argillaceous rocks in the upper part. There are also a few amphibolites, derived from gabbroic intrusives. Three of the sedimentary formations deserve special mention because they comprise most of the rocks in the Homestake mine. The oldest formation, the Poorman formation, is graphitic, ankeritic dolomite and ankerite rock with minor amounts of phyllite and recrystallized chert bands. The next younger formation, the Homestake formation, is sideroplesite schist with many bands of recrystallized chert; in part of the district the sideroplesite schist has been regionally metamorphosed to cummingtonite schist. The next younger formation, the Ellison formation, is mainly phyllite with many beds of graphitic quartzite. The reader is referred to earlier papers (Dodge, 1942; Noble and Harder, 1948) for full descriptions. The schists were folded isoclinically in pre-Cambrian time and were further deformed by the intrusion of porphyry stocks and dikes in Eocene time (Noble, Harder, and Slaughter, 1949).

Structural Control of Mineralization

Ore Bodies

The briefest characterization that can be made of the Homestake ore bodies is that they are selective replacements of the sedimentary Homestake formation. Practically every ton of ore produced has come from that formation, and the small amount obtained from the adjacent Poorman formation would not have made a profitable mining operation. Only certain parts of the Homestake formation have been mineralized to ore grade, however, and the selective character of the process of mineralization is structural.

A belt less than a mile wide covers all the known ore of significant amount in the Homestake mine, and in this belt five distinct ore bodies have been mined. From east to west, these are: the Caledonia; the main ore body; the no. 5 ledge; the east no. 9 ledge; and the west no. 9 ledge. Plate 1 shows part of the structure on the 2300 level. Here the Caledonia ore body is a narrow, attenuated syncline near the northeast edge of the mine area. The Homestake formation continues beyond the Caledonia to the northeast as a very narrow bed which on the surface becomes highly deformed as it approaches a porphyry stock. Westward from the Caledonia syncline the Homestake formation continues in somewhat uneven but more normal thickness through several folds, with one interruption by faulting, until it reaches the complex structure of the main ore body, which is developed on the structure of the Pierce anticline. Westward from the main ore body the formation is interrupted and deformed by a zone of rhyolite dikes, and beyond the dikes is the synclinal structure of the no. 5 ledge. This syncline correlates with the De Smet syncline on the surface. From the no. 5



COMPOSITE MAP OF THE NO. 5

The levels are separated to avoid superposition. O

COMPOSITE MAP OF THE NO. 5 LEDGE ORE BODY

The levels are separated to avoid superposition. Ore is shown by solid black.

LEDGE ORE BODY

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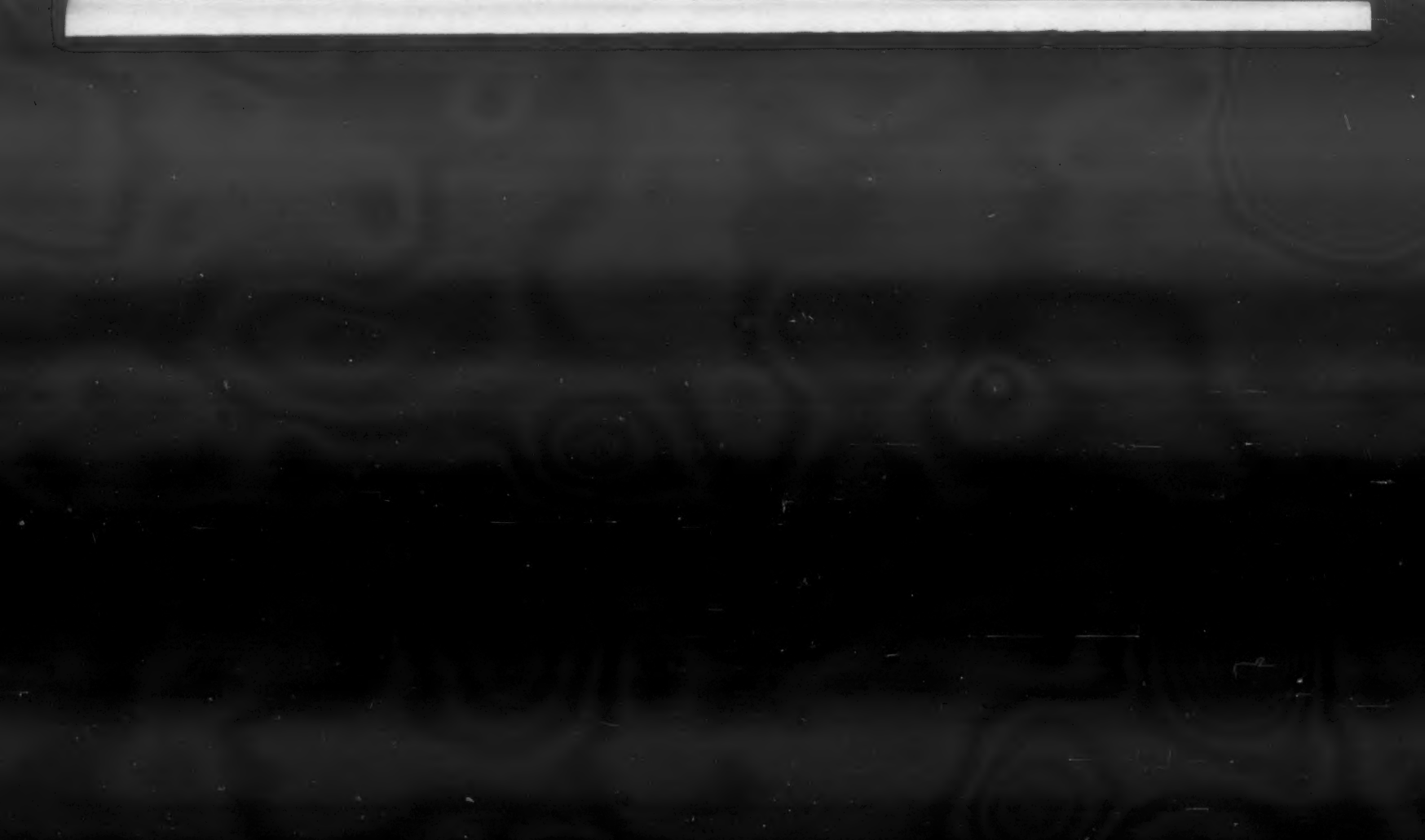


ledge ore body the Homestake formation passes through the Independence anticline, an unnamed syncline, the no. 9 ledge anticline, the Lead syncline in which are developed the east and west ore bodies of the no. 9 ledge, and thence to the no. 11 ledge anticline.

These several synclines and anticlines are isoclinal folds dipping steeply east and plunging southeast. Presumably these folds were developed largely by shear folding, although some transverse compression and bending may have been important in the earlier stages of the folding. The largest structure on the level shown in Plate 1 is the Independence anticline. The other named folds and many others not named are subsidiary to this structure. The whole area of Plate 1 is subsidiary to the much larger Poorman anticline, which dominates the structure of the pre-Cambrian rocks in the Lead district (Noble, Harder, and Slaughter, 1949). The Poorman anticline lies west of the mine area, beyond the limits of Plate 1.

In order of commercial importance, the five ore bodies are the Main ore body, West no. 9 ledge, East no. 9 ledge, Caledonia, and no. 5 ledge. There are also a few areas in which enough mineralization has been found to justify some stoping operations, but none of these areas has yet shown real continuity.

Viewed broadly, each ore body is a long, doubly tapering spindle-shaped pipe or pencil pointed down moderately steeply to the southeast. This essentially linear or pipelike shape is clear in the smaller ore bodies. It fits the main ore body only with some generalization in the upper levels, where that ore body is very large, but is an accurate characterization in the lower levels where the ore body is smaller.

Localization in Cross Folds

Linear or pipelike ore bodies can be formed in a number of ways, and for these shapes in the Homestake mine several explanations have been offered. However, one explanation is now favored to explain all five principal ore bodies. This is that those ore bodies are localized in zones of a second-stage folding—that is, cross folds superimposed on the earlier isoclinal folds. The angle of intersection between the axial planes of the isoclinal folds and the cross folds is only 20° or 30°. Planar features, the shear planes of the second stage of folding, thus intersect other planar features, either the isoclinal axial planes or one limb of isoclinal folds, and the resulting intersections make the linear features of the ore bodies. Here again the smaller ore bodies best fit the hypothesis, the larger ore bodies only with some generalization. The details of shape and interpretation of the mechanism are discussed for each ore body.

The no. 5 ledge ore body, the smallest of the named ore bodies, illustrates most concisely the localization of ore in cross folds, because this ore body is restricted to the interval between the 1850 and the 3050 levels and has been completely explored. In Plate 2 the outlines of the structure and ore areas are shown to scale and in orientation but offset to avoid overlap. The structure plunges southeast rather steeply. The highest recognizable structure which can be called a part of the no. 5 ledge structure is a small bulge on the side of the De Smet syncline on the 1850 level. On the next level below (2000 level), the ore-bearing structure is considerably larger, and the cross folding extends across the axial plane of the syncline. On the 2150 level there is little change, but below the 2150 level the syncline is notably folded, and the area of the structure increases progressively to the 2600 level; thereafter it decreases rapidly almost to the vanishing point on the 3050 level. No corresponding structure is known either above the 1850 level or below the 3050 level.

The area of commercial ore shows a similar progressive increase in depth to a maximum and then a rapid decrease to the vanishing point. The maximum ore area is on the 2750 level (Table 1).

The Caledonia structure and ore body, though not so well known as the no. 5 ledge ore body, is cut by the erosion surface near the top of the structure and is explored nearly to the bottom in the mine workings. Plate 3 is a composite map to scale but with each level offset from the next to avoid overlap. The structure plunges southeast, steeply in the upper levels and flatter in the lower levels. On the surface (on Pl. 3 projected to the zero reference level) the structure is a partly closed syncline, moderately cross folded. A progressive increase of cross folding is accompanied by a marked increase in area through each level to the 400 level, where the synclinal portion of the fold

becomes a separate pod. Thenceforth there is a progressive decrease in area to the 700 level but with even more acute cross folding. On the 800 level the synclinal structure becomes much simpler and shows only a small amount of cross folding, and this condition continues to the lowest level developed, the 2300 level. The maximum area of ore is on the 500 level, and the maximum percentage of ore is on the 900 level. The areas of structure, areas of ore, and percentages of ore on each level are given in Table 2.

TABLE 1.—AREA OF HOMESTAKE FORMATION AND OF ORE IN THE NO. 5 LEDGE STRUCTURE

Level	Area of Homestake formation sq. ft.	Area of ore sq. ft.	Percentage of ore in formation
1700	none	none	0
1850	1,500	75	5.0
2000	4,550	1,000	22.0
2150	3,400	1,925	56.6
2300	8,300	2,300	27.7
2450	14,400	4,650	32.3
2600	22,700	5,300	23.3
2750	16,350	10,250	62.7
2900	5,600	2,725	48.7
3050	750	none	0
3200	none	none	0

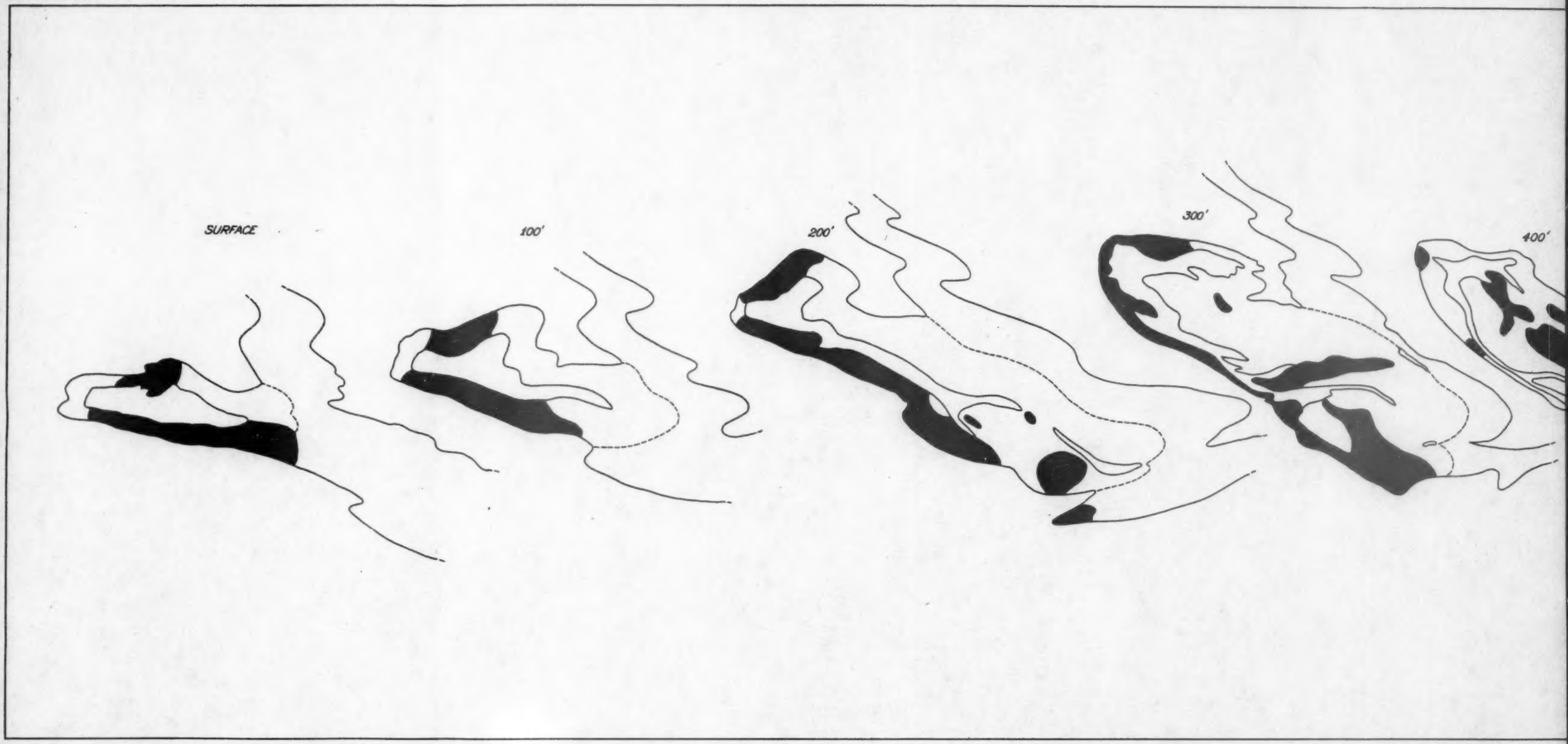
The structure which contains the ore body known as the main ore body is undoubtedly the most complicated of the mine area, and study of this structure led to the concept of the structural control of ore in the cross folds. Because the area of the ore-bearing folds is quite large on the surface and still continues to the deepest mine level (the 5000 level), no attempt is made to show the changes from level to level except between the 2300 and 2900 levels. The cross folding of the main ore body has been discussed (Noble, Harder, and Slaughter, 1949). Figure 1 shows the deformations superimposed on the axial plane of the Pierce anticline, and the localization of ore for the same levels is shown in Figure 2. Four important cross folds and one minor one are shown on these figures, and the positions of the cross folds correspond well with the distribution of the ore shoots. In the upper levels of the mine, especially above the 900 level, the individual cross folds become widely dispersed, whereas in the lower levels of the mine, below the 3350 level, they either combine or some disappear. The net result for the ore-bearing structure is a gradual reduction in cross-section area from the surface to the lowest levels, but for short distances the change may not be marked or may even be in the opposite direction, as for example in the interval shown in Figures 1 and 2.

TABLE 2.—AREAS OF HOMESTAKE FORMATION AND OF ORE IN THE CALEDONIA STRUCTURE

Level	Area of Homestake formation sq. ft.	Area of ore sq. ft.	Percentage of ore in formation
zero	40,000	not known	not known
100	126,000	not known	not known
200	236,000	not known	not known
300	264,500	21,275	8.0
400	247,300	58,850	23.8
500	228,400	83,300	36.5
600	175,400	60,700	34.6
700	116,700	51,975	44.5
800	42,625	21,925	51.4
900	34,575	21,275	61.5
1000	31,275	14,800	47.3
1100	27,475	10,400	37.9
1250	19,600	5,800	29.6
1400	12,500	4,550	36.4
1550	10,700	4,050	37.9
2300	11,400	1,900	16.5

Figure 3, a cross section of the main ore body, shows the nature of the deformation in a vertical section.

The three ore bodies described are localized in cross folds which deform axial planes of synclines in the no. 5 and the Caledonia ore bodies and an anticline in the main ore body. The structures in which are localized the east and west ore bodies of the no. 9 ledge, on the other hand, are cross folds which involve only one limb of the isoclinal structure, the large Lead syncline. The deformations caused by the cross folds are in fact so similar to minor folds related to the major syncline that the concept of cross folding is adopted mainly by analogy to the cross-folded structures previously described. Features indicative of cross folding are U-shaped folds with cleavage parallel to bedding



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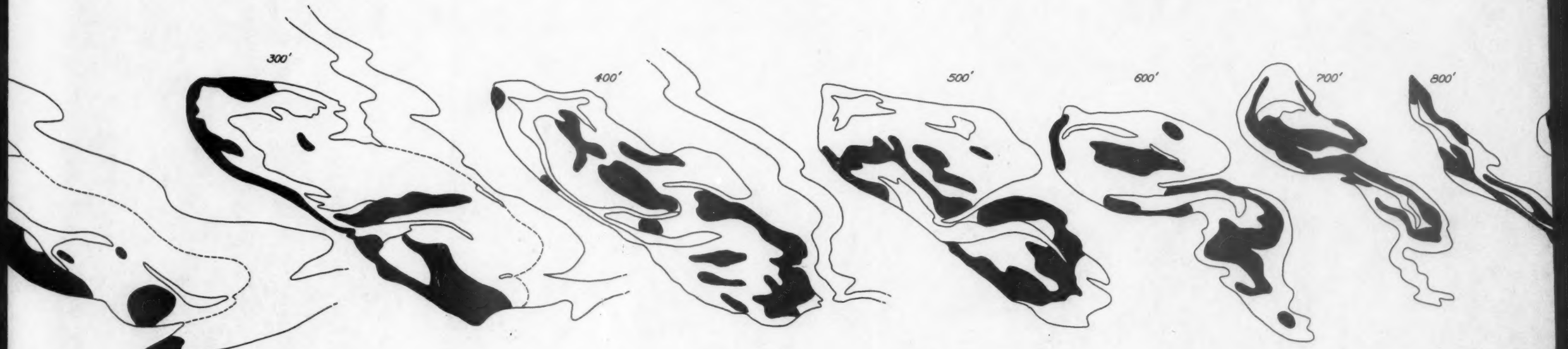
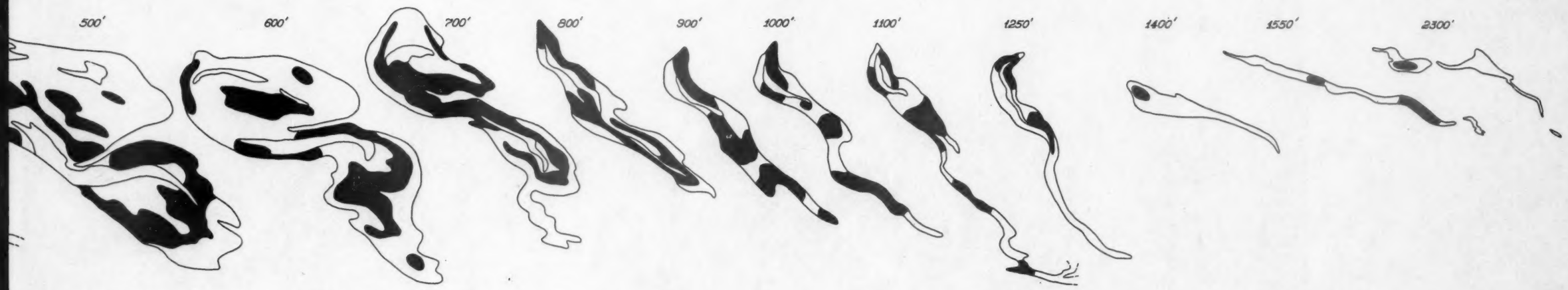


COMPOSITE MAP OF THE CALEDONIA ORE BODY

The levels are separated to avoid superposition. Ore is shown by solid black.



LEDONIA ORE BODY

Ore is shown by solid black.

around the folds, and appreciable thickening of the Homestake formation.

The two no. 9 ledge ore bodies (east ore body and west ore body) as developed from the 2300 level to the 3200 level are shown in Plate 4, a composite map in which the levels are separated to avoid overlap. These structures plunge southeast. Within the interval shown, the east ore body is a spindle-shaped ore body localized in a minor fold on the east side of the syncline. This minor fold is interpreted as a cross fold which intersects the east limb of the syncline at a low angle, causing a displacement of west side south or up.

The folds in which are localized the west ore body of the no. 9 ledge are also interpreted as cross folds which displace the east limb of the Lead syncline. The axial plane of the Lead syncline is the most westerly syncline shown on Plate 4. The shearing in the cross folds nearly parallels the strike of the axial plane, and the displacement is west side south or up. Two or three closely related cross folds are believed to make up the structure. Ore mineralization tends to be localized in the separate zones but to a considerable degree crosses from one zone to the next. No place is known, however, where ore crosses from the east ore body to the west ore body, presumably because the band of unsheared rock between the two bodies was wide enough to resist penetration by mineralizing solutions.

Dilatation as a Cause of Ore Localization

Small bodies of quartz show the same localization in small cross folds that is shown by the ore bodies in the larger cross folds. In Figure 1 of Plate 5 three quartz veins are localized in small cross folds, and Figure 2 of Plate 5 is a closer view of one of these folds with its accompanying quartz vein. The closely folded rocks have been refolded by small discontinuous zones of cross folding, and this refolding has been attended by a considerable dilatation or increase in volume of the rock in the zones of cross folding, shown by a thickening of individual beds. Quartz has then replaced part of each zone of cross folding. In the vein on the right of Figure 1 of Plate 5, a considerable part of the dilated zone has been replaced, so it might be argued that here the injection of quartz displaced the vein walls. However, in the smaller vein on the left of Figure 1 of Plate 5, shown to larger scale in Figure 2 of Plate 5, the quartz has clearly replaced the schist rather than been injected.

The principle of dilatancy is certainly not new in geology (Mead, 1925), but it has rarely been used to explain structural control of ore mineralization (Roberts, 1941). As applied to the Homestake ore bodies, the theory is that the nearly perfect dimensional orientation of mineral grains in the isoclinally folded rocks has been disturbed by fracturing and rotation of grains in the zones of cross folding, thereby increasing the volume of the rock by increasing the intergranular porosity.

The first evidence in support of this theory of ore localization by dilatancy should perhaps come from microscopic study, but this line of approach is not wholly satisfactory because the deformed rock is generally replaced almost completely by hydrothermal minerals. In the vicinity of the ore bodies shown in Plates 2, 3, and 4, the rock minerals are extensively replaced by chlorite, sericite, and carbonates. This is especially noticeable where the original rock is largely cummingtonite, because the early stage of mechanical destruction of cummingtonite can be readily seen, and in places the direction of shearing can be related to the direction of displacement in the cross folds. Fractured cummingtonite is not so common as chloritized areas in cummingtonite, however, and probably most of the fractured cummingtonite was replaced by chlorite.

Crenulations in schists are caused by reorientation of minerals in tiny zones of cross folding, and, insofar as this orientation is by bending and granulation, the crenulated zones must be dilated. Plate 6 shows the extent to which the mineral grains are thrown out of orientation in crenulations. This is probably the smallest scale in which dilatation is important, as these crenulations are the width of only two or three mineral grains.

There are probably all sizes of cross folds from that of the crenulations to that of the Homestake ore bodies or even larger structures. A size above that of the crenulations is fairly common in the Homestake mine (Pl. 5). Since the plane of the crosscut wall shown in Plate 5

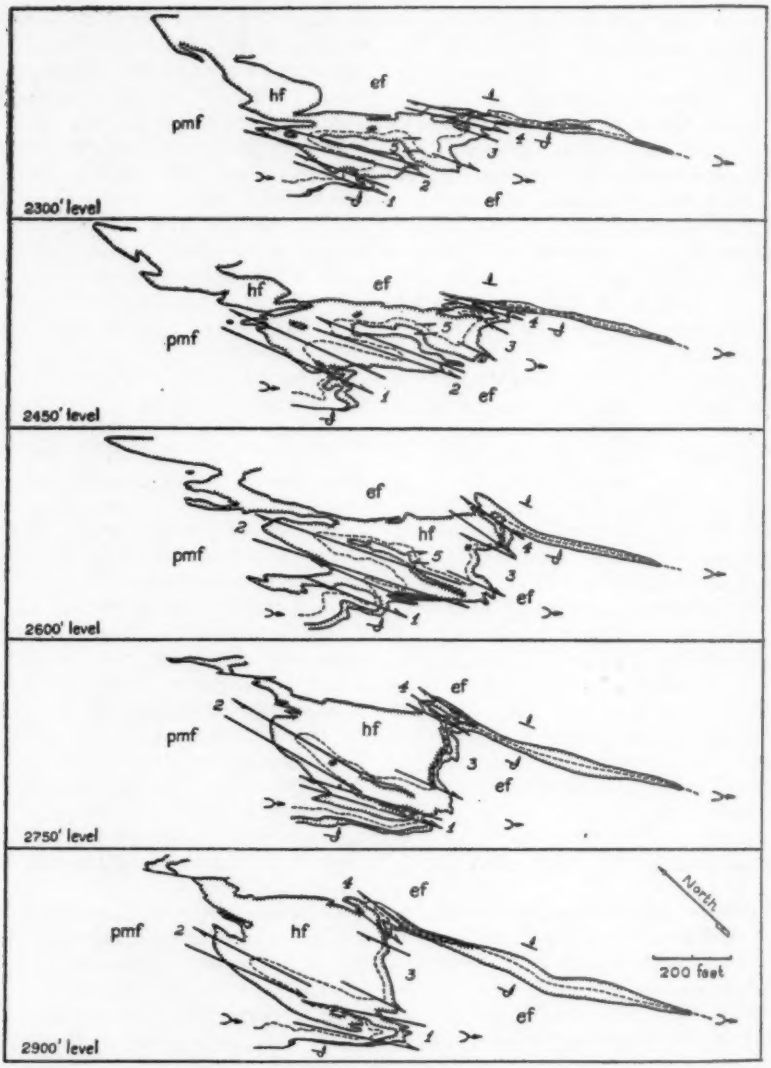


FIGURE 1.—DEFORMATION OF THE PIERCE ANTICLINE
Cross folds shown by shear arrows. *See* Figure 2.

is at right angles to the strike of the strata, changes in thickness measure changes in volume of rock. Thickening of strata in the zone of cross folding is clearly shown.

There are many cross folds of various scales between these and the folds of the ore bodies. Many are seen on the walls of stopes, but they are rarely sufficiently well exposed to be photographed. Folds of several different sizes are included in the folds of the ore bodies. Of these, probably the structure of the no. 5 ore body shows the effects of dilatation most clearly (Pl. 2). The change in area on the horizontal sections of the different mine levels (Tables 1,

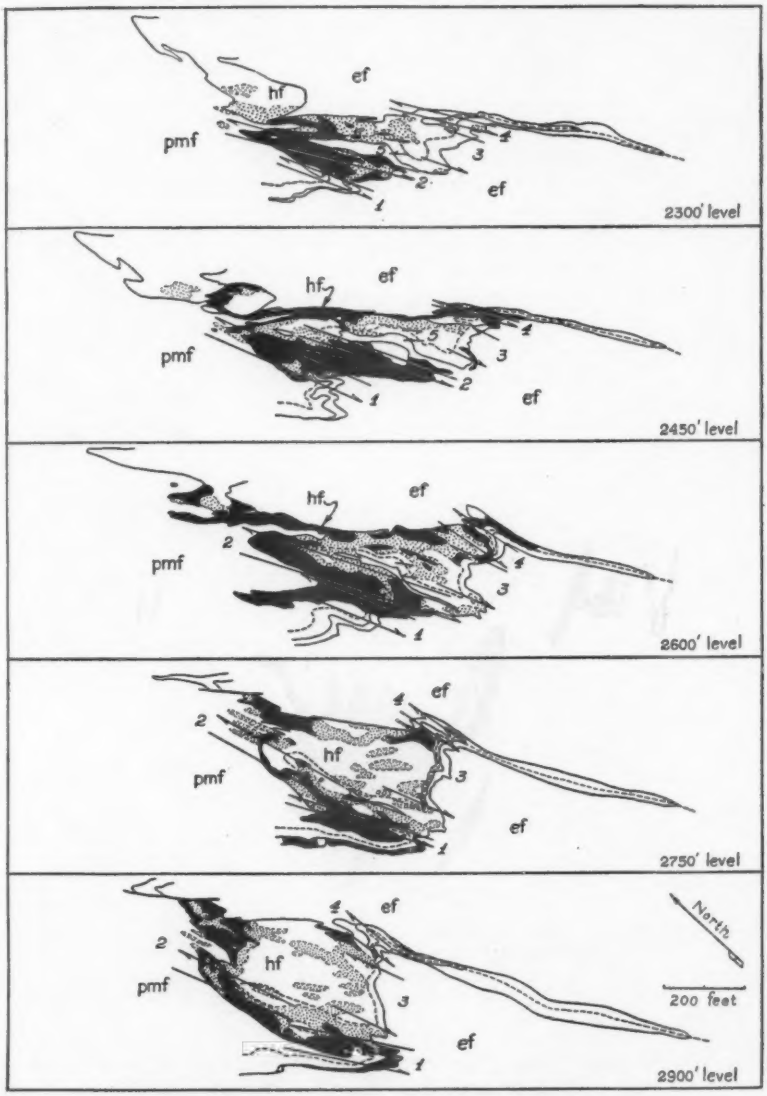


FIGURE 2.—LOCALIZATION OF ORE (SOLID BLACK) IN THE CROSS FOLDS OF THE PIERCE ANTICLINE

See FIGURE 1.

2) is due in part to dilatation and in part to repetition by minor folding or crumpling. The Caledonia structure (Pl. 3) shows about the same amount of change of area, due in part to dilatation. The deformation of the folds which form the main ore body is complicated by proximity to the rhyolite dikes, but on most levels it has clearly resulted in an increase in area as compared with adjacent parts of the Homestake formation. On Figures 1 and 2 there

FIGURE 3.—VERTICAL CROSS SECTION OF THE MAIN ORE BODY

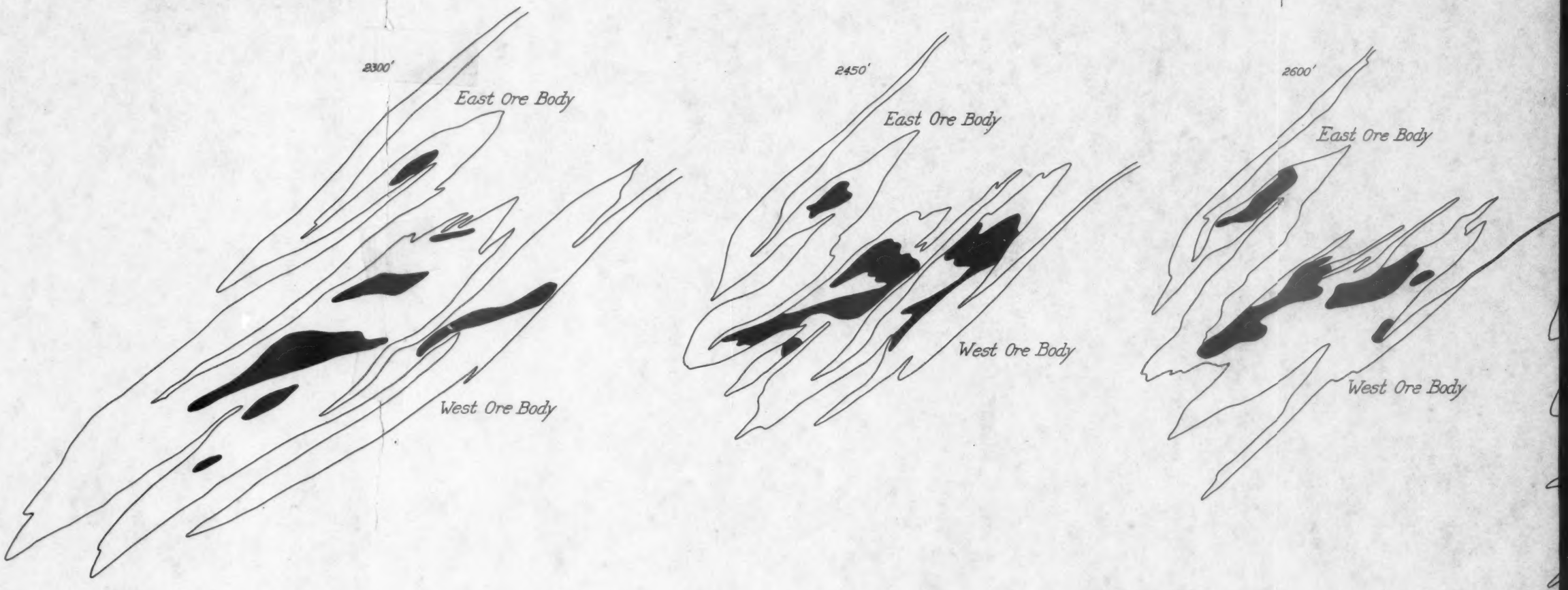
The isoclinal axis of the Pierce anticline is strongly deformed by a cross fold, and the ore (solid black) is localized in the zone of cross folding. Symbols: pmf, Poorman formation; hf, Homestake formation; ef, Ellison formation.

is an appreciable increase in area on the lower levels (2750 and 2900). The structures of the east and west ore bodies of the no. 9 ledge ore bodies (Pl. 4) are appreciably larger than adjacent parts of the Homestake formation, presumably due in part to dilatation.

Lack of Localization by Faults, Anticlines, or Dikes

The cross folds which have localized the ore bodies of the Homestake mine are not equivalent to faults and do not pass into faults in the overlying formations. A few definite faults of small displacement do occur in the Homestake mine, and a large fault along the east edge of the district has probably been crossed in the lower levels of the mine, but neither the large fault nor the smaller ones have any noticeable effect in localizing ore.

Anticlinal features have been cited as a structural control of ore in many places (Newhouse, 1931), and at times an anticlinal control has



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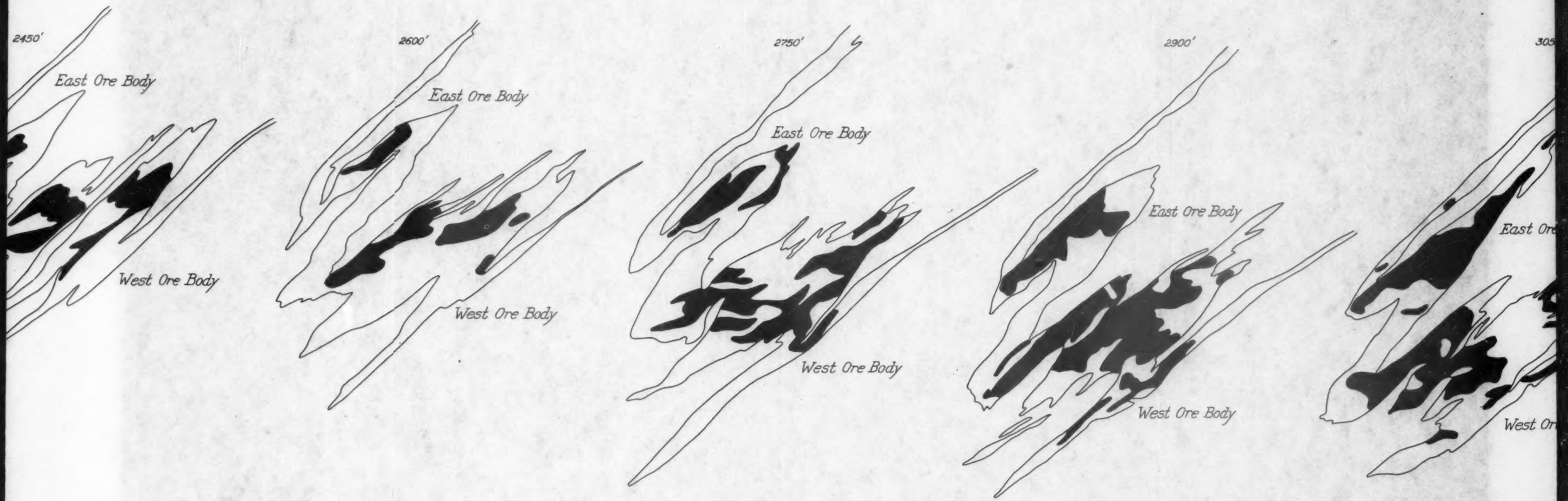
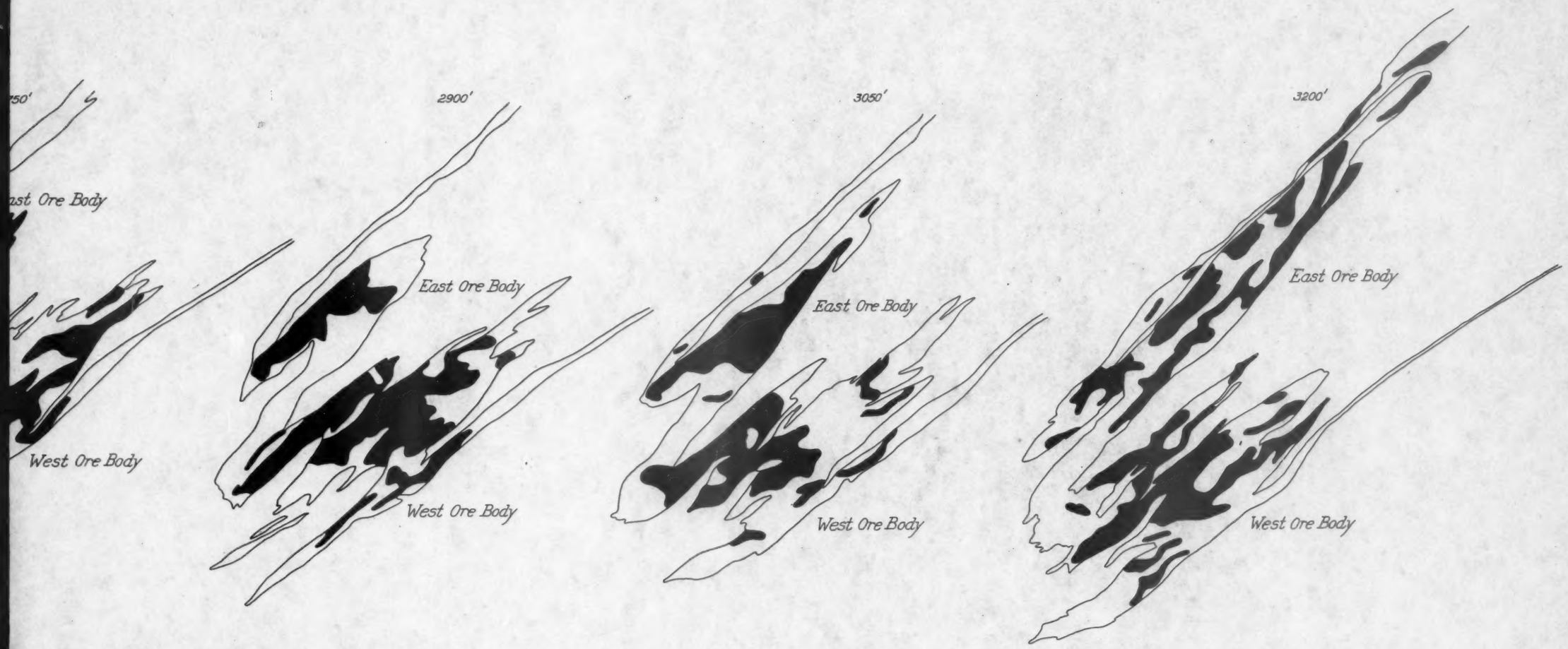
The levels are separated



COMPOSITE MAP OF PART OF NO. 9 LEDGE ORE BODIES

The levels are separated to avoid superposition. Ore is shown by solid black.



O. 9 LEDGE ORE BODIES

Ore is shown by solid black.

appeared to be important in the Homestake mine; but by the new concept of localization of ore in cross folds it is not the anticlinal shape but the shape of the cross fold which is important, and probably there is no important effect of concentration of flow of solutions under an impermeable arch. This is shown by the fact that where ore appears to be under an impermeable cover the overlying formations were in fact penetrated by hydrothermal solutions, so that they carry much quartz and some other minerals but lack other minerals, notably gold. An additional weakness of the anticlinal theory of control is that in the simple cross folds ore extends from the anticline to the adjacent syncline (*see* the east ore body of no. 9 ledge in Pl. 4), and in the synclinal folds (no. 5 ledge and Caledonia, Pls. 2, 3) there is no anticlinal control. On the other hand, in a few places there are weak concentrations of mineralization in the crests of isoclinal folds such as the Pierce, Independence, and no. 11 ledge anticlines. The distribution of ore does not strongly suggest control by the arch of the fold, and it seems more probable that this control may also be dilatation due to rotation of grains during the shear folding which produced the isoclinal folds (Noble, Harder, and Slaughter, 1949, p. 326). The theory of structural control which related the Homestake ore bodies to the rhyolite dikes could be supported only so long as the ore bodies and dikes were together. That relationship holds for the main ore body from the surface to about the 2450 level, but below that level the dikes and ore body gradually separate, and on the 5000 level the dikes and ore body are far apart. None of the other ore bodies shows any close relationship to the dike zone.

The theory of structural control of ore by zones of cross folding, therefore, explains all the principal ore bodies of the Homestake mine. Not all cross folds are ore-bearing, however; some cross folds as large as these ore-bearing folds have been fairly thoroughly explored without disclosing mineable ore. Other factors than the simple mechanism of dilatation in cross folds must also be important. The large cross folds, like the small crenulations, presumably are nonpersistent, and the flow of mineralizing solutions had to be adjusted to channels of permeability which are probably too complicated to explain. At the present stage in the development of the theory of structural control of the Homestake ore bodies the best exploration policy seems to be to follow the Homestake formation into zones of cross folding rather than to follow zones of cross folding to intersections with the Homestake.

Stages and Sequence of Mineralization

Field Relations

The ore bodies of the Homestake mine are more or less completely chloritized portions of the cummingtonite or sideroplesite schist of the Homestake formation, with fairly abundant veins and masses of quartz and a sprinkling of pyrrhotite, pyrite, and arsenopyrite. For an ore in which gold is the only valuable constituent, the amount of sulphides is unusually high. The total sulphides average between 7 and 8 per cent in the mill feed, distributed in proportions of about 50 per cent pyrrhotite, 40 per cent pyrite, and 10 per cent arsenopyrite. Minor hydrothermal constituents are ankerite, cummingtonite, biotite, garnet, albite, calcite, sericite, fluorite, galena, sphalerite, chalcopyrite, specularite, magnetite, gypsum, and gold. Although this large assortment of hydrothermal minerals occurs in intimate association and may be found within a single stope, the association is not haphazard. Study over a considerable area shows a definite sequence of stages of mineralization, especially for the ore of the main ore body. Failure of the no. 9 ledge ore to conform will be discussed later.

The Homestake formation, before the arrival of the hydrothermal solutions of the gold mineralization, had been highly folded, converted in part from sideroplesite schist to cummingtonite schist, and refolded by one or more periods of cross folding. In parts of the mine area untouched by the hydrothermal mineralization the formation consists of sideroplesite or cummingtonite, biotite, quartz, and minor amounts of garnet and graphite. Layers and pods of dark-gray recrystallized quartz, mostly highly contorted, are fairly abundant. These were probably originally chert.

Study of the field relations enables separation of the hydrothermal mineralization in the Homestake mine into four distinct stages. The minerals of the first stage are quartz, chlorite,

and arsenopyrite, formed by replacing the minerals of the schists. Quartz probably was the first to form, and the earliest stage was a replacement of the layers and pods of dark-gray recrystallized quartz. In a number of places there is preserved an incipient replacement of the dark-gray quartz by white hydrothermal quartz which was guided by tiny parallel fractures. More commonly the replacement was nearly complete, and adjacent veinlets merged to form large masses of hydrothermal quartz about parallel to the schist structure. Only the incipient stage of introduction of quartz was confined to the bands of dark-gray recrystallized quartz. From these first replacement veins the quartz masses continued to grow by replacement of the schist minerals, but in the main the veins preserved a general conformity to the schist structure (Pl. 7, fig. 1). Accompanying or closely following the introduction of the quartz, chlorite extensively replaced the cummingtonite schist. Cummingtonite disappeared from the walls of most of these conformable quartz veins, and in the areas of most intense mineralization the cummingtonite schist over widths of scores of feet was completely converted to a dense greenish-black rock which is essentially chlorite and hydrothermally recrystallized biotite and quartz. The areas of chloritized rock, parallel to the conformable veins of quartz, are about parallel to the schist structure. Arsenopyrite is abundant in the chloritized rock along the walls of many of the conformable veins of quartz and is disseminated erratically throughout the cummingtonite schist. The euhedral arsenopyrite crystals range from 2 or 3 mm to about 2 cm in diameter. In places they are closely crowded into groups or lines of grains, especially around the edges of masses of quartz, and in other places they are widely scattered.

Fracturing seems to have played no part in the distribution of the minerals of this first stage of mineralization, except for the small amount of fracturing in the bands of dark-gray recrystallized quartz. The hydrothermal quartz was guided by the tiny bands of recrystallized quartz, the chlorite was guided by the masses of hydrothermal quartz, and the arsenopyrite was disseminated throughout the chloritized rock.

The second stage of mineralization consists of quartz (continuing from the first stage), ankerite, pyrrhotite, and minor amounts of albite, biotite, garnet, and cummingtonite, the last three probably merely transported and redeposited from the schist walls. Straight, sharp-walled, crosscutting quartz veins are moderately abundant in cummingtonite schist adjacent to chloritized areas, and most of these veins carry minor amounts of ankerite or pyrrhotite or both. The veins run in many directions, most commonly probably about north, at about 35° from the schist structure. Opposite walls of these veins have not been faulted. The intersection of bedding planes with the veins shows that these veins, although guided by fractures, are wholly replacements, not open-space fillings. Cross-cutting veins of this type are virtually absent from chloritized rock, probably because the chloritized rock was too weak to form continuous fractures. This relationship and the fact that in the outer borders of the chloritized areas conformable quartz-chlorite-arsenopyrite veins are cut by quartz-ankerite-pyrrhotite veins prove the difference in age of the first and second stages of mineralization. Ankerite and pyrrhotite were also deposited throughout the areas of chloritized rock. Ankerite occurs in

PLATE 5.—VEIN QUARTZ LOCALIZED IN SMALL CROSS FOLDS

FIGURE 1.—THREE QUARTZ VEINS LOCALIZED IN SMALL CROSS FOLDS

FIGURE 2.—CLOSER VIEW OF ONE OF THESE VEINS

Showing dilatation of beds and replacement by vein quartz in zone of cross folding. Scale is 6 inches long.

PLATE 6.—PHOTOMICROGRAPHS OF CRENULATED MICA SCHIST

FIGURE 1.—MICA SCHIST CUT BY TINY CROSS FOLDS WHICH CAUSE MINOR DISPLACEMENT OF BEDDING PLANES

Plane polarized light.

FIGURE 2.—SAME AS FIGURE 1, CROSSED NICOLS

In the cross folds micas are bent and thrown out of orientation. Both figures cover a standard thin section, ×3.

FIGURE 1

FIGURE 2

VEIN QUARTZ LOCALIZED IN SMALL CROSS FOLDS

FIGURE 1

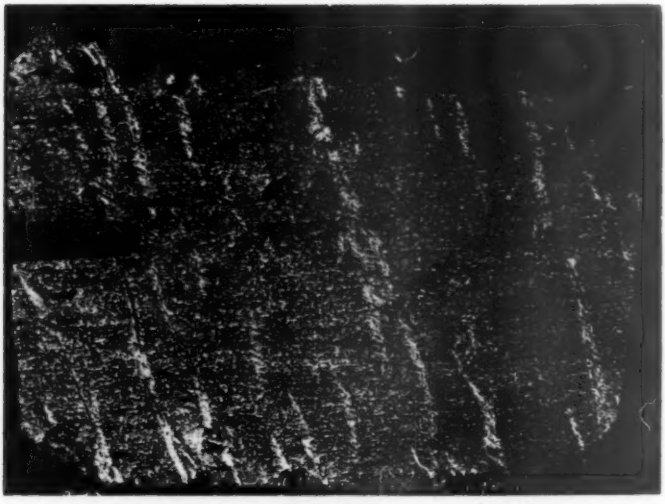


FIGURE 2

PHOTOMICROGRAPHS OF CRENULATED MICA SCHIST

FIGURE 1

FIGURE 2

VEIN QUARTZ REPLACING FOLDED SCHIST

FIGURE 1

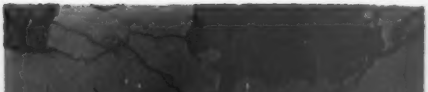


FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

FIGURE 6

FIGURE 7

FIGURE 8

PHOTOMICROGRAPHS OF POLISHED SECTIONS OF HOMESTAKE ORES

tiny irregular seams and in patches, in very erratic distribution. Pyrrhotite occurs in tiny specks and elongated groups of grains throughout practically all the chloritized rock and in many places in cummingtonite schist. It is the most abundant sulphide in the Homestake ores.

The sizes of individual quartz masses and the relative abundance of quartz veins differ greatly from place to place. On some levels, notably on the 2450 level, there are very large masses of conformable quartz of the first stage of mineralization, and on other levels there are very large crosscutting masses of quartz of the second stage of mineralization. Stoping operations show that these large masses have no great vertical persistence. The size or relative abundance of these large quartz masses is unrelated to the extent or amount of other minerals accompanying the quartz. On the other hand, good ore seldom lacks abundant small quartz veins.

Albite, biotite, garnet, and cummingtonite are in or adjacent to a few small quartz-ankerite veins. Some tiny veins contain quartz, ankerite, and garnet; others contain quartz, ankerite, and biotite; and others contain quartz, biotite, and garnet. Most of the albite is in small, very irregular masses of apple-green anhedra. Biotite occurs in shiny black plates up to 1 cm in diameter, typically in or just along the walls of small quartz veins. The garnet occurs in or along the walls of tiny quartz veins and scattered throughout cummingtonite schist. The euhedral crystals range from 2 mm to about 3 cm in diameter, and some large irregular masses are several centimeters in diameter. This garnet is dark red. Hydrothermal cummingtonite, as distinguished from the rock cummingtonite produced by regional metamorphism, is very rare in the Homestake mine. When present it is commonly asbestiform and very pale brown. Tiny blades of cummingtonite penetrate the edges of some small quartz veins, and some

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PLATE 7.—VEIN QUARTZ REPLACING FOLDED SCHIST

FIGURE 1.—QUARTZ AND ACCOMPANYING MINERALS REPLACE FOLDED HOMESTAKE FORMATION ALONG BEDDING PLANES

FIGURE 2.—REPLACEMENT VEINS OF QUARTZ CROSS CONTACT BETWEEN HOMESTAKE FORMATION (left and lower part) AND ELLISON QUARTZITE (upper right part)

Contact was marked by a chalk line before the photograph was taken. High grade ore in Homestake formation; no gold values in Ellison formation.

PLATE 8.—PHOTOMICROGRAPHS OF POLISHED SECTIONS OF HOMESTAKE ORES

FIGURE 1.—PYRRHOTITE (GRAY) FORMING BORDER AROUND EUHEDRAL CRYSTAL OF ARSENOPYRITE (WHITE)

Pyrrhotite is later than arsenopyrite. ×35.

FIGURE 2.—PYRRHOTITE (GRAY) VEINING A CRYSTAL OF ARSENOPYRITE (WHITE)

Pyrrhotite is later than arsenopyrite. Also pyrrhotite inclusions in arsenopyrite. ×35.

FIGURE 3.—PYRITE (GRAY) REPLACING EUHEDRAL CRYSTAL OF ARSENOPYRITE (WHITE), REMNANTS OF WHICH ARE IN PARALLEL OPTICAL ORIENTATION

Pyrite is later than arsenopyrite. ×35.

FIGURE 4.—EUHEDRAL CUBES OF PYRITE (WHITE) REPLACING LENTICULAR AREA OF PYRRHOTITE GRAINS (GRAY) WITHOUT MODIFYING LENTICULAR SHAPE

Pyrite is later than pyrrhotite. ×45.

FIGURE 5.—SPECULARITE (GRAY) REPLACING PYRRHOTITE (WHITE) ALONG PARTING DIRECTIONS

Specularite is later than pyrrhotite. ×45.

FIGURE 6.—MAGNETITE (GRAY) IN PLATES PSEUDOMORPHOUS AFTER SPECULARITE WHICH HAD REPLACED PYRITE (WHITE)

One pyrite grain is cut into pieces by the plates; hence specularite is later than pyrite, and magnetite is later than specularite. ×35.

FIGURE 7.—MARCASITE (SMALL WHITE GRAINS) ALONG EDGE OF GANGUE VEINLET WHICH CUTS PYRRHOTITE (LIGHT GRAY)

Marcasite is later than pyrrhotite. ×35.

FIGURE 8.—GOLD (WHITE) REPLACING PLATES OF SPECULARITE (LIGHT GRAY)

Some lines of residual grains of specularite are in parallel optical orientation; hence gold is later than specularite. ×70.

hydrothermal garnet and albite are replaced by cummingtonite.

In general, in the main ore body most quartz-chlorite-arsenopyrite veins are conformable to the schist structure and are surrounded by chloritized rock, and most crosscutting veins in cummingtonite schist beyond the limits of chloritized rock belong to the later stages of mineralization. However, a few crosscutting quartz veins accompanied by chlorite and arsenopyrite reach out into the cummingtonite schist for many feet. Chlorite replaces the cummingtonite schist for distances of a few inches or a few feet along the walls of these veins. Moreover, a few of the quartz-ankerite and quartz-pyrrhotite veins contain a little chlorite and arsenopyrite as well. In most of the ore of the main ore body the first period of well-defined fracturing separates the quartz-chlorite-arsenopyrite stage from the quartz-ankerite-pyrrhotite stage, but in some places chlorite and arsenopyrite continued in minor amounts into the later stage.

The third stage of mineralization consists of pyrrhotite alone, continuing from the second stage. Tiny straight veins only 1–2 inches wide but many feet long cut the cummingtonite schist and both the earlier stages of mineralization. The walls of most of these veins have been faulted a few inches or 1–2 feet. Most of these veins contain only massive pyrrhotite, but a few of them contain small partly replaced grains of quartz, and in a very few there are partly replaced grains of arsenopyrite. Probably a little quartz and arsenopyrite continued to form after the beginning of the fracturing of this third stage of mineralization but were largely replaced by the pyrrhotite. The field relations permit no doubt that as a whole the veins of massive pyrrhotite are later than both the earlier stages of mineralization. The veins of massive pyrrhotite are not abundant.

The fourth and last stage of mineralization is separated from the earlier stages by the intrusion of the Tertiary rhyolite dikes and by much fracturing and some faulting. The rhyolite dikes clearly cut and brecciate the quartz-chlorite-arsenopyrite and the quartz-ankerite-pyrrhotite mineralizations. That the dikes have not been seen to cut the veins of massive pyrrhotite is probably due merely to the rarity of the veins. This fourth stage of mineralization consists mainly of pyrite and calcite, with very minor amounts of quartz, sericite, fluorite, celestite, anhydrite, gypsum, chlorite, opal, dolomite, rhodochrosite, specularite, magnetite, sphalerite, galena, chalcopyrite, realgar, and native arsenic. Narrow, straight or gently curving veinlets of calcite and pyrite, most of them somewhat vuggy and some with large open spaces, are abundant throughout the east half of the mine area. In the lower levels they are most common in two zones, one near the rhyolite dikes (some cutting the dikes) and the other along the east side of the mine near a zone of faulting. Many minor faults have been partly filled by calcite and pyrite. Along many of these veinlets there has been no alteration of the wall rocks, but along some the wall rocks for many feet are extensively replaced by pyrite, calcite, sericite, and in a few places a little magnetite and specularite. Pyrite is not confined, however, to the vicinity of the calcite veinlets but is abundant in many parts of the earlier products of mineralization. Within a few feet of the rhyolite dikes and near some calcite veins pyrite has completely replaced other sulphides. Specularite and magnetite occur along the walls of some dikes and within the dikes. Veins of calcite, sphalerite, galena, chalcopyrite, and in a few places fluorite are found in the schists, especially near the dikes, and they cut the dikes in a few places. A few tiny veins of gypsum of unquestionable hypogene origin cut the Poorman formation. In a small vein of calcite with a lttle sphalerite on the 3800 level, gypsum filled vugs and was the last mineral to form. Veinlets of gypsum have been noted as deep as the 4700 level. On the 3500 level a small calcite vein cutting a quartz vein carried fluorite, native arsenic, realgar, and gypsum.

Some large crevices lined by coarse calcite and a few crystals of pyrite are probably late phases of this fourth stage of mineralization. Most of these crevices when encountered by mine workings are filled with water strongly charged with H_2S. One crevice cut by a long diamond drill hole on the 2300 level, at that time the bottom of the mine, was filled with water containing the following constituents in parts per million (N. Herz, analyst):

Na	428
K	60
Ca	13
Mg	11
SiO_2	11
$Al_2O_3 + Fe_2O_3$	4
CO_3	392
SO_4	351
S_2O_3	20
H_2S	15
Cl	4
Total	1309

It seems improbable, from the position at which this crevice was encountered in the drill hole and the proportions of dissolved constituents, that this water could have been contaminated appreciably by surface waters. Soda is practically absent from the schists and the surface waters in the Lead district. Therefore the water and dissolved constituents probably are end phases of the hydrothermal mineralization. Lindgren (1933, p. 53), discussing waters of this composition, states that "the prevailing opinion is that these waters with their charge are in whole or in part of magmatic origin".

A few grains of pyrrhotite have been noted in some calcite-pyrite veinlets. The pyrrhotite must have persisted from the earlier stage of mineralization. Similarly in a few calcite veinlets a very little ankerite probably is carried over from the preceding stage. Moreover, a few calcite veins are accompanied by chlorite (daphnite).

The veins of the calcite-pyrite stage of mineralization strike in almost every direction. Many strike about north; others strike about N. 30° W., which is a direction of minor faulting along the east side of the lower levels of the mine. Some veins split to make two or three veins, but others intersect without joining. Some calcite-pyrite veins follow the walls of the earlier veins of massive pyrrhotite. Clearly, at the time of the pyrite-calcite stage of mineralization the Homestake formation had become quite thoroughly fractured.

In a few places in the outer parts of the ore zone adjacent fractures of similar attitude have guided mineralizations of distinct stages. Veins of massive pyrrhotite in a few places parallel veins of the quartz-ankerite-pyrrhotite stage, and in other places veins of the calcite-pyrite stage parallel those of massive pyrrhotite or of quartz, ankerite, and pyrrhotite. Some crosscutting veins of the quartz-chlorite-arsenopyrite stage parallel veins of the quartz-ankerite-pyrrhotite stage.

The four stages of mineralization recognized in the main ore body are less distinct and not easily separated in the no. 9 ledge ore bodies, mainly because the stages are not separated by distinct structural breaks. Some fracturing had already developed before the introduction of the first stage of mineralization, and the intersecting relations of the stages are somewhat confused. In addition some of the age relations are contradictory, and there may have been overlaps or repetitions of stages of mineralization in these no. 9 ledge ore bodies. Crosscutting quartz-chlorite-arsenopyrite veins are common, and many of them have very flat dips.

The position of the gold of the Homestake mineralization in relation to the four stages of mineralization cannot be determined by the field studies. Within the limits of the main ore body there is an obvious association of gold with the quartz-chlorite-arsenopyrite mineralization. Not only is most of the visible gold found in chloritized rock immediately along the walls of quartz veins, but much of the invisible gold which makes most of the ore accompanies arsenopyrite. Massive vein quartz does not carry gold, however. Every Homestake miner knows that in the main part of the mine the presence of arsenopyrite guarantees that the rock is of ore grade. In addition much rock without arsenopyrite also carries gold. Nearly all the chlorite-pyrrhotite schist which surrounds the conformable quartz masses is good ore, and also within the ore zone much of the unchloritized cummingtonite schist carries gold. On the other hand, in a few outlying parts of the mine there is no close relation between gold and arsenopyrite, and some specimens with abundant arsenopyrite and chlorite carry almost no gold.

Gold is found by assay but not by eye in some cummingtonite schist bordering the zones of chloritized rock. This cummingtonite schist has almost no chlorite and no arsenopyrite but does have some fine-grained pyrrhotite. Possibly the gold has moved beyond the area of the quartz-chlorite-arsenopyrite mineralization, but more

probably this gold was introduced during or after the quartz-ankerite-pyrrhotite stage of mineralization. Gold in small amounts (to the value of about 0.15 oz. per ton) is also found in the narrow veins of massive pyrrhotite, and this gold probably is at least as late as the third stage of mineralization. Finally, some gold is found in the calcite-pyrite stage of mineralization. Metallurgical tests show that clean concentrates of pyrite carry appreciable amounts of gold (Wayland, 1934), and assays show that pyrite-calcite veins in barren cummingtonite schist carry gold.

The field relations indicate, therefore, that gold has been introduced either during each stage of mineralization, or during the last stage of mineralization and has replaced the products of the earlier stages of mineralization.

Differences in amounts of minerals in different parts of the Homestake mine constitute a sort of mineral zoning. (1) Pyrrhotite virtually disappears in the upper levels of the mine, as noted by Sharwood (1911, p. 756), and indicated by Irving and Emmons (Irving, Emmons, and Jaggar, 1904, p. 68, 82–92), who found only one doubtful piece of pyrrhotite in a suite of specimens from levels down to and including the 800. Although the upper levels of the main ore body are not now accessible, the upper levels of the Caledonia ore body show almost no visible pyrrhotite above the 400 level (about 600 feet below the surface). There is, however, microscopic pyrrhotite in some specimens which show no visible pyrrhotite. An increase in pyrite accompanies the decrease of pyrrhotite in most of these upper-level ores, but in a few specimens the only sulphide is arsenopyrite. In some specimens coarse pyrite crystals line vugs in vein quartz. (2) Coarse visible gold is more abundant in the upper levels and in the no. 9 ledge ore bodies than in the lower levels of the main ore body. A fairly large percentage of specimens of good ore selected at random from the upper levels and from the no. 9 ledge ore bodies shows visible gold in hand specimens or under the microscope, whereas few specimens of good ore from the lower levels of the main ore body show visible gold. (3) Pyrite is least abundant in the west ledges. (4) Chalcopyrite, though only microscopic and nowhere abundant, is in much greater amount in upper-level ores than in those from the lower levels.

On the other hand, the late, vuggy calcite-pyrite veins of the fourth stage of mineralization probably show no relation to the present erosion surface. Although the distribution of these veins is erratic, they seem to be as abundant in the lower mine levels as in the upper levels.

These mineralization products formed in the Homestake formation. Mineralization was not confined to that formation, however, and one can trace the course of the mineralizing solutions in the Poorman and Ellison formations. Mineralization of the quartz-chlorite-arsenopyrite stage extends into the Poorman formation for a few feet or even for a few scores of feet in some zones of strong cross folding, and hydrothermal chloritization of schist minerals extends into the Poorman formation for hundreds of feet from the ore zones in the Homestake formation. The only obvious difference from the mineralization of the same stage in the Homestake formation is that the quartz veins and masses are somewhat smaller, and chloritization is incomplete. The quartz-ankerite-pyrrhotite mineralization also extends into the Poorman formation in some places, but the veins here are typically conformable rather than crosscutting. In addition to quartz-ankerite-pyrrhotite veins there is a very large amount of fine-grained pyrrhotite in the Poorman formation, and this pyrrhotite extends for many scores, perhaps hundreds, of feet from the zones of mineralization in the Homestake formation. Quartz veins unaccompanied by other hydrothermal minerals are quite abundant in some parts of the Poorman formation. The narrow, straight veins of massive pyrrhotite of the third stage of the Homestake mineralization have not been noted in the Poorman formation. Calcite-pyrite veins are, however, probably as abundant in the Poorman formation as in the Homestake formation. The zones of cross folding which guided the ore solutions produced shearing and disturbances in the Poorman and Ellison formations as well as in the Homestake formation, and there is no reason to believe the solutions moved more readily through the Homestake formation than in the adjacent formations. Some of the products of mineralization do not stop at the contact between Home-

stake and Poorman formations, although only a small amount of arsenopyrite and gold are found across that contact.

In the Ellison formation, on the other hand, almost no hydrothermal minerals were deposited except quartz, calcite, and pyrite. Only a few grains of arsenopyrite have been noted, and pyrrhotite is rare. Ankerite is found in only a few places. Chlorite accompanies quartz across the contact in some places, and some of the phyllites of the Ellison formation are slightly chloritized for a few scores of feet from the zones of mineralization in the Homestake formation. Typically the minerals which accompany quartz stop at the contact between Homestake and Ellison formations, even where the quartz veins continue across into the Ellison (Pl. 7, fig. 2). A few of the narrow, straight veins of massive pyrrhotite cross from Homestake formation into Ellison, and these change, within a few feet along the strike, to calcite-pyrite veins. Calcite-pyrite veins cross without change from the Homestake into the Ellison formation and are probably as abundant in one formation as in the other.

Although quartz veins are relatively scarce in the phyllites of the Ellison formation, they are very abundant in the quartzites. Most of the quartz veins in the quartzites are straight, sharp-walled, parallel replacement veins. In some places they follow directions which are probably shear directions produced by cross folding. Some of this quartz is clearly a part of one of the recognized stages of mineralization of the Homestake ore bodies (Pl. 7, fig. 2). On the other hand, vein quartz in straight, sharp-walled, parallel replacement veins is found in most outcrops of Ellison quartzite in the northern Black Hills, even many miles from Lead. Since it seems improbable that the mineralizing solutions which produced the Homestake ore bodies travelled so widely, probably the Ellison quartzite was readily converted to vein quartz by many different hydrothermal solutions.

The chemistry of the replacement of quartzite by quartz is not easily explained. Figure 2 of Plate 7 shows that the process is one of replacement, and that the same veins formed by replacement of cummingtonite schist in the Homestake formation and of quartzite in the Ellison formation. The white vein quartz is pure SiO_2, and the dark quartzite is SiO_2 plus a very small amount of graphite. There is no visible concentration of graphite adjacent to the quartz veins, as a result of crowding out of graphite by recrystallization of the quartz.

Microscopic Study of the Ores

The field relations seen in mine exposures and those deduced from a study of the mine maps can be supplemented by microscopic study. Field relations show a somewhat overlapping succession of mineralization products: quartz; chlorite; arsenopyrite; ankerite with subordinate albite, garnet, biotite, and cummingtonite; pyrrhotite; pyrite and calcite with subordinate sphalerite, galena, chalcopyrite, and fluorite; and gold. The microscopic study of thin and polished sections of mineralized specimens supports the results of field investigations and also discloses several additional minerals and gives more precise age relations. Unfortunately, microscopic relations are subject to different interpretations by different observers, and it is therefore desirable to discuss the criteria used in determining the age relations.

Two criteria of age relations of nonopaque minerals to other nonopaque minerals and to opaque minerals in the Homestake ores have been used: pseudomorphism and intersection. Idiomorphism has not been used because intersections are abundant and permit more positive interpretations. Pseudomorphous replacements most common in the Homestake ores are biotite by chlorite and cummingtonite by calcite. The most common intersections are cummingtonite and biotite cut by sulphides and sulphides and quartz veined by sericite and calcite.

The age relations of opaque minerals are much more difficult to interpret. The opaque minerals here, as in most ore deposits, have been introduced rather suddenly and erratically from a deep-seated source, and they formed mainly by replacing rock minerals, early-forming gangue minerals, or other opaque minerals. The process of deposition of opaque minerals in an ore deposit has little in common with the process of crystallization of minerals in a magma, and the student of polished sections of ore specimens can adopt criteria of

age relations applicable to polished sections but not to thin sections of igneous rocks (Ross, 1935, p. 14). Furthermore, where a single process of mineralization has operated, as the process of replacement in the Homestake ore body, it may be possible to formulate criteria applicable to that process but not to another process, for instance that of open-space filling.

In a study of approximately 250 polished sections of Homestake ore specimens a great many relationships of one mineral to another were described and tabulated. Although probably most of the relationships have some significance, only a relatively small number of them give age relationships with reasonable certainty. Table 3 lists those geometrical relations of one opaque mineral to another which seem significant in determining age relations in the Homestake specimens. Throughout this study an attempt was made to find criteria indicating contemporaneous formation of minerals, since contemporaneity or overlap of products of mineralization most probably occurred, but no good criteria of contemporaneity were found. These criteria therefore fix the end of the period of deposition of a mineral but do not fix the beginning. The question of the amount of overlap and of the possibility of repetition of minerals is also discussed.

The microscopic study of the polished sections subdivides the opaque minerals into three groups on the basis of age and association. The first group comprises arsenopyrite, pyrrhotite, and pyrite; these are found in nearly every specimen of typical ore in the Homestake mine. Arsenopyrite is the oldest and in many sections had been somewhat fractured before pyrrhotite was introduced. Pyrite is definitely later than pyrrhotite. There is no proof of more than a slight amount of overlapping or contemporaneous deposition of arsenopyrite, pyrrhotite, or pyrite. A second group of minerals, later than the first, comprises specularite, magnetite, and marcasite. Small amounts of specularite and magnetite are found in most sections, but marcasite is less widely distributed. Magnetite is later than specularite, but the relation of marcasite to the other two was not determined. Marcasite can be placed in any one of the three groups, since it is later than arsenopyrite, pyrrhotite, and pyrite but has not been related in age to any other opaque mineral. It is included with specularite and magnetite because it is believed that these minerals form in similar chemical environments. On the basis of the field relations these three minerals in the Homestake mine can only be hypogene. The third group of minerals is later than the arsenopyrite-pyrrhotite-pyrite group but could not be fixed in relation to the specularite-magnetite-marcasite group. It comprises tetrahedrite, galena, sphalerite, and chalcopyrite. All are found only in small amounts and in a few specimens, except chalcopyrite, which is found in small amounts in nearly all specimens. The age relations are difficult to establish, but a few convincing relations give a sequence in the order listed. Gold was seen in about one-fourth of the polished sections. It shows associations with the arsenopyrite-pyrrhotite-pyrite group and also with the tetrahedrite-galena-sphalerite-chalcopyrite group; the most definite associations are with arsenopyrite and with galena. Gold definitely is later than arsenopyrite, pyrrhotite, pyrite, magnetite, specularite, and galena.

Arsenopyrite and pyrrhotite are older than the rhyolite dikes, because they are strongly fractured and replaced by pyrite along the walls of the dikes, but all the other minerals are younger than the dikes, in part at least.

Because pyrite is very abundant in the Homestake ores and because much of the pyrite is definitely younger than the Tertiary rhyolite dikes, the question of more than one generation of pyrite calls for special study and discussion. Most pyrite in the Homestake sections is euhedral or subhedral and has relatively few or no gangue inclusions. In a few sections, however, there is another type having very irregular form, always anhedral, and commonly full of tiny gangue inclusions. In a very few instances the anhedral type of pyrite was molded around or veined the euhedral type and so was later. No other mineral intervenes between the two kinds of pyrite, and probably the second type is separated from the first by a very short time interval. Since the earlier euhedral type is abundant in and alongside the rhyolite dikes, there is no possibility of making one pyrite pre-Cambrian in age and the other Tertiary.

In the absence of acceptable criteria for con-

TABLE 3.—SPACE RELATIONSHIPS IN POLISHED SECTIONS OF HOMESTAKE ORE

Significant Relationships

Veining and localization in fractures

1. True subsequent veins.
   Examples: pyrrhotite veining arsenopyrite; pyrite veining arsenopyrite; gold in cracks in arsenopyrite.
2. Penetrations in irregular, somewhat veinlike areas.
   Example: pyrrhotite in arsenopyrite.
3. Metallic grains in gangue veinlets cutting earlier sulphides.
   Examples: marcasite in gangue veinlet in pyrrhotite; gold in carbonate veinlet in arsenopyrite.
4. Sulphide localized along wall of a gangue veinlet cutting earlier sulphide.
   Examples: chalcopyrite replacing pyrrhotite along a gangue veinlet; marcasite replacing pyrrhotite and pyrite.

Peripheral relationships

5. Small grains perched on faces of larger grains; usually the larger grain is euhedral.
   Examples: pyrrhotite on arsenopyrite; pyrite on arsenopyrite; chalcopyrite on pyrite.
6. Earlier larger grain replaced around the edges by smaller grains.
   Examples: pyrite replacing pyrrhotite; specularite replacing pyrrhotite and pyrite; marcasite replacing pyrrhotite and pyrite.
7. Euhedral inclusions in a larger grain surrounded by a border of smaller grains.
   Example: euhedral pyrite in pyrrhotite surrounded by chalcopyrite which replaces the pyrrhotite; hence chalcopyrite is later than both pyrrhotite and pyrite.
8. Replacement remnants in one mineral surrounded by small grains of another mineral.
   Example: arsenopyrite remnants in pyrrhotite surrounded by chalcopyrite which replaces the pyrrhotite; hence chalcopyrite is later than both arsenopyrite and pyrrhotite.

Miscellaneous

9. Localization in subsequent features of an earlier mineral.
   Examples: marcasite in parting planes of pyrrhotite; pyrrhotite in cleavage planes of albite; specularite in parting direction of pyrrhotite.
10. Fracturing attendant on crystallization.
    Example: euhedral pyrite grain in pyrrhotite surrounded by radiating fractures (rare). May be localization in a pre-existing fracture zone, in which case it is covered by example no. 1.

Relationships Generally Significant but Open to Misinterpretation

Pseudomorphism

11. Unit pseudomorphs.
    Example: magnetite after specularite (may not be units).
12. Aggregate pseudomorphs.
    Example: aggregates of pyrite after units of arsenopyrite.
13. "Pseudomorphism" of larger features.
    Example: pyrite grains selectively replacing lenticular areas of pyrrhotite in schist; pyrite is euhedral against pyrrhotite but is anhedral against schist and preserves the original lenticular outline.

Peripheral relationships

14. One mineral between grains of another mineral.
    Examples: pyrrhotite between grains of arsenopyrite; galena between grains of arsenopyrite. These suggest that the pyrrhotite and galena are later, but antecendent veins show the same relationsip with a different conclusion.

Penetrations

15. Plates or blades of one mineral penetrating or cutting another mineral.
    Example: specularite plates penetrating or cutting across grains of pyrrhotite or pyrite. Suggests that specularite is later, but molding of one grain about another gives the same relationship with a different conclusion.

TABLE 3.—*Continued*

Inclusions

16. Optically parallel inclusions of one mineral in another (the penultimate stage of the "exploded bomb" texture, which is almost unknown here).
Examples: arsenopyrite inclusions in pyrrhotite; pyrrhotite inclusions in pyrite and in specularite; specularite in gold. Usually it is safe to conclude that the inclusions are residual; but note that skeletal crystals will have the same relationship but a different conclusion.

Relationships not Significant

Euhedrism

17. Rounded or ragged inclusions of minerals generally euhedral.
Examples: rounded inclusions of arsenopyrite in pyrrhotite, of quartz in pyrrhotite. Since arsenopyrite and quartz are generally euhedral against pyrrhotite, lack of euhedrism suggests replacement by pyrrhotite.
18. Ragged edges to grains usually euhedral.
Example: ragged grains of arsenopyrite in gangue. Since arsenopyrite is generally euhedral, lack of euhedrism suggests replacement by gangue.

Inclusions

19. Small inclusions of one mineral in large units of another.
Examples: pyrrhotite in arsenopyrite; chalcopyrite in pyrrhotite and in sphalerite; gold in arsenopyrite and in pyrrhotite. No conclusion can be drawn from this relationship. But if the inclusions do not occur in the adjacent rock, they are not earlier than the host mineral; they may be either contemporaneous or later.
20. Small inclusions of one mineral in inclusions in another mineral.
Examples: gold in inclusions of pyrrhotite in arsenopyrite; chalcopyrite in inclusions of pyrite in pyrrhotite. No conclusion can be drawn, unless the small inclusions are localized on contacts, in which case they are probably later than the other minerals.
21. Dimensionally oriented inclusions.
Examples: specularite in arsenopyrite and in pyrite (rare). Significance is uncertain. Suggests burial on a growing face, hence specularite overlapped by the other minerals; but in at least one section the specularite plates crossed grain boundaries. May be a habit of specularite.

temporaneity or overlap it was possible to fix only the time a mineral ceased to form, not that at which it began to form. The relations between pyrite and the two oxides of iron, specularite and magnetite, may not permit the detection of overlap. In many polished sections the oxides are later than pyrite and have formed by alteration or replacement of it. Possibly the stage of oxide formation was brief, however, and sulphide deposition was resumed very soon, but the oxides were not replaced or altered by the sulphide-bearing solutions. On the other hand, the microscopic study probably rules out the possibility of recurrence of other minerals in the Homestake ores. The order of deposition of the first group of opaque minerals is arsenopyrite, pyrrhotite, pyrite. In this sequence pyrite is not in its normal place; in most similar deposits it comes first (Schwartz, 1937a, p. 33). If there is a second generation of pyrite, it followed directly after the first, because all the minerals of the two later groups are later than pyrite. In the same way each mineral of the Homestake ores occupies a single place in a table of paragenesis; if a mineral has formed in two generations, the second has followed the first with no new mineral between them. For instance, chalcopyrite occurs with the group of relatively rare minerals tetrahedrite, galena, and sphalerite, but it also occurs in very small amount in nearly all specimens. It is most common in pyrrhotite, especially surrounding small cubes of pyrite enclosed in pyrrhotite. If this widely distributed chalcopyrite associated with pyrrhotite and pyrite is an earlier generation nearly contemporaneous with the pyrrhotite and pyrite, then probably some of the later minerals would be later than some chalcopyrite, but such relations have not been seen. The most probable interpretation is that chalcopyrite, though forming a very late stage in the mineralization, spread

widely throughout the mineralized zone. The same reasoning applies to the question of the age of the gold. Much of the gold is later than arsenopyrite and veins or replaces it. Some is later than pyrrhotite, pyrite, and magnetite. Some gold is later than galena. If the gold which replaces arsenopyrite is an earlier generation, probably some of that earlier gold would have been veined or partly replaced by some of the later minerals. The writer prefers the interpretation that the gold, though forming at the end of the mineralization, has spread throughout the early-formed products of mineralization and in one place or another has replaced nearly every earlier mineral.

Microscopic Descriptions of the Opaque Minerals

ARSENOPYRITE: Euhedral and subhedral grains of arsenopyrite from 1 mm to 2 cm in size are typical. In a very few specimens the arsenopyrite is in ragged anhedra. Arsenopyrite replaces many schist minerals and is replaced and veined by all the other opaque minerals and by quartz, calcite, and sericite. In most polished sections a few grains of arsenopyrite are fractured.

PYRRHOTITE: The most abundant sulphide is pyrrhotite. It is found in all pre-rhyolite specimens except some from the upper levels of the main ore body. It typically occurs as irregular anhedra in schist. Some areas are rounded, and others are lenses elongated parallel to the schistosity. In a few specimens pyrrhotite forms lines of grains or tiny veinlets. A very few subhedral prisms or plates of pyrrhotite in schist have been seen. Small clusters of pyrrhotite grains surrounding arsenopyrite and areas which vein and replace arsenopyrite clearly show that pyrrhotite is later than arsenopyrite. The rather common occurrence of tiny irregular inclusions of pyrrhotite in arsenopyrite is more difficult to interpret. These inclusions probably show replacement of arsenopyrite by pyrrhotite because (1) the proportion of the two minerals is very erratic from grain to grain and within single grains of arsenopyrite; many grains have no inclusions; and (2) there are all gradations from a wholly random pattern of tiny inclusions to large veinlike replacements. Pyrrhotite shows little evidence of shearing or fracturing, but most of the sheared pyrrhotite may have been replaced by other minerals, especially pyrite. Pyrrhotite has replaced nonopaque minerals, notably quartz, chlorite, garnet, ankerite, and albite, but it is also replaced by some nonopaque minerals, especially calcite and sericite.

PYRITE: Pyrite, second only to pyrrhotite in abundance except in the west ledges, is the most abundant opaque mineral in the ore near the rhyolite dikes and on some of the upper levels. Typically it forms lines or groups of euhedral or subhedral grains in schist. It also occurs as scattered cubes and irregular areas of anhedra. Another form, slightly later than these, is in irregular areas crowded with tiny irregular inclusions of gangue. Finally, a form possibly still later is that of tiny branching veinlets of anhedra. Groups of tiny grains of pyrite surround euhedra of arsenopyrite, and ragged areas of pyrite replace arsenopyrite. Some rhombic areas of tiny grains of pyrite are aggregate pseudomorphs of pyrite after arsenopyrite, and some of these show remnants of unreplaced arsenopyrite. Irregular and subhedral inclusions of pyrite occur in some grains of arsenopyrite. Replacement of pyrrhotite by pyrite is common, and much of the pyrite is euhedral. In many sections pyrrhotite in schist has been nearly completely replaced by pyrite, but the tiny inclusions of pyrrhotite in arsenopyrite remain. Possibly under some conditions arsenopyrite was impermeable to solutions depositing pyrite, or perhaps pyrrhotite outside arsenopyrite was strained by shearing and therefore was susceptible to conversion to pyrite whereas that enclosed in arsenopyrite was protected. Considerable pyrite of the Homestake mine is moderately anisotropic, and it has been suggested (Schneiderhohn and Ramdohr, 1931, p. 159; Schwartz, 1937b, p. 818) that anisotropic pyrite contains arsenic; but in the Homestake ores much of the anisotropic pyrite has replaced pyrrhotite or nonopaque minerals and shows no relation to arsenopyrite. Some pyrite is fractured, and a little is veined by sericite and calcite.

Five specimens of Homestake ores do not support the general order of disposition. (1) A grain of pyrite was cut by a veinlet of pyrrhotite. (2) An "exploded bomb" of pyrite was partly replaced by pyrrhotite. (3) A veinlet of pyrite cut a grain of magnetite. (4) In a specimen from a quartz vein bearing visible gold in the Ellison formation (a type of mineralization that has been noted in only three or four places over a period of several years), pyrite is earlier than specularite but is also later than sphalerite. (5) In a specimen from one of the few veins of massive pyrrhotite in the Ellison formation pyrite is both earlier and later than pyrrhotite. The writer interprets these specimens as showing that momentarily and locally there was repetition of conditions which governed the deposition of these different minerals.

SPECULARITE: Strongly anisotropic, highly reflecting specularite is widespread in small amounts

in all kinds of ores from the Homestake mine. It is commonly in scattered plates or irregular grains in schist, mostly under 1 mm in diameter but in two specimens 3 or 4 mm in diameter. There are some rosettes of plates and a few tiny veinlets. Some lines of plates wrap around grains of arsenopyrite. Specularite replaces arsenopyrite, pyrrhotite, and pyrite and is included in those minerals. Parallel lines of platy inclusions of specularite occur in arsenopyrite and in pyrite. In two specimens, gold cuts and replaces coarse plates of specularite.

MAGNETITE: Magnetite, probably as abundant as specularite and found in all kinds of ores from the Homestake mine, is most abundant in the pyritic replacements in the walls of rhyolite dikes and calcite veins. Most of it is in irregular grains in schist or in gangue minerals. A few grains are euhedral. Some magnetite replaces arsenopyrite, pyrrhotite, and pyrite irregularly, but it is most common as pseudomorphous replacements of specularite.

MARCASITE: Marcasite has been detected in very small amounts in a few specimens of all types of ore from the Homestake mine. Most commonly it is in tiny grains or veinlets in pyrite, especially replacing pyrite around the edges. In specimens which have two ages of pyrite marcasite replaces the later pyrite. Pyrrhotite is replaced by marcasite but not so extensively as is pyrite. Sharp-pointed parallel veinlets of marcasite in pyrrhotite grains may be replacements on parting planes. A few residual inclusions of arsenopyrite in pyrrhotite are surrounded by tiny grains of marcasite. Marcasite is therefore later than arsenopyrite, pyrrhotite, and pyrite, but its age has not been determined in relation to the other opaque minerals. It replaces calcite in some tiny veins in pyrrhotite. In a few specimens, tiny veinlets of marcasite cut schist, and in some of these there is also a little pyrite.

TETRAHEDRITE: Only a few grains of tetrahedrite have been detected in Homestake specimens. One specimen is schist at the contact with a rhyolite dike, and two specimens are fillings of calcite veins. The grains of tetrahedrite are intimately associated with galena, sphalerite, and chalcopyrite. In one polished section a veinlet containing tetrahedrite, galena, and sphalerite cuts a grain of pyrite.

GALENA: Galena in very small amounts is fairly common in the specimens where mineralization accompanies or is later than the rhyolite dikes but is very rare in the normal specimens. The most common occurrence is in small irregular areas or veinlets in schist and in gangue minerals. Euhedral grains are very rare. Galena borders and replaces arsenopyrite and replaces both pyrrhotite and pyrite.

SPHALERITE: Sphalerite, like galena, is found in very small amounts in much of the ore associated with the rhyolite dikes but very rarely in the normal ore. It occurs as tiny grains in schist and gangue minerals and replaces both pyrrhotite and pyrite, especially around the edges. In one specimen it definitely replaces galena, and in another specimen it seems to replace tetrahedrite.

CHALCOPYRITE: Although chalcopyrite belongs with tetrahedrite, galena, and sphalerite in age and everywhere accompanies those minerals, it is also widespread in small amounts in all but about 25 specimens. It is somewhat more abundant in the upper levels than in the lower. It occurs in part in grains and veinlets in schist and in gangue minerals but much more commonly replaces other opaque minerals, particularly pyrrhotite. Some patterns of tiny grains of chalcopyrite in pyrrhotite might be interpreted as a result of exsolution, but there is so much evidence of veining and replacement that probably the inclusions of chalcopyrite are also replacements. Pyrite is veined and replaced by chalcopyrite, and euhedral grains of pyrite in pyrrhotite are surrounded by chalcopyrite. Some grains of sphalerite have an "emulsion" texture of tiny grains of included chalcopyrite, which might be interpreted as a result of exsolution; but other grains of sphalerite are veined and replaced by chalcopyrite, and probably the "emulsion" texture here is a result of replacement. Some grains of galena are surrounded and replaced by chalcopyrite.

GOLD: Gold, like chalcopyrite, occurs in irregular grains and veinlets in schist and in gangue minerals but more commonly replaces other opaque minerals. Probably more gold is found in arsenopyrite than in any other opaque mineral, but, instead of being directly enclosed in the arsenopyrite, most of it has partly replaced small inclusions of pyrrhotite, pyrite, or gangue in the arsenopyrite. Gold also fills fractures in arsenopyrite and is perched on euhedral faces of arsenopyrite grains. Both pyrrhotite and pyrite contain small inclusions of gold, some of which are clearly replacements by gold. In a few specimens magnetite is replaced by gold, in some chalcopyrite is replaced by gold, and in others galena is replaced by gold. In two specimens there are grains of gold in calcite veins. Sharwood (1911, p. 754) had noted a similar occurrence of gold, fluorite, and a bismuth telluride in a vug in a calcite vein.

Photomicrographs of polished sections of Homestake ores illustrate the occurrences of the minerals and the criteria of age relations (Pl. 8).

Microscopic Descriptions of the Nonopaque Minerals

QUARTZ: Replacement masses of quartz form coarse, flamboyant grains and also aggregates of small grains with sutured or mosaic boundaries. Most of the tiny veinlets filling fractures in opaque minerals are calcite veinlets, but a few also have some quartz and perhaps chalcedony. Quartz cuts and replaces the schist minerals and in a few places cuts arsenopyrite, but arsenopyrite is later than most of the vein quartz.

CHLORITE: Chlorite is widely distributed throughout the Homestake ores. Gustafson (1930; 1933) and Dodge (1935; 1942) determined the chlorite of the Homestake ores to be daphnite, with index γ between 1.625 and 1.652. It has a conspicuous ultra-blue interference color. Gustafson distinguished between hydrothermal chlorite and "rock" chlorite (*i.e.*, a product of regional metamorphism), but recent studies suggest that all the chlorite of the Homestake ores is hydrothermal. The chlorite replaces the cummingtonite, sideroplesite, biotite, and quartz of the schists extensively and with other minerals replaces garnet. Chlorite also cuts and replaces hydrothermal quartz, garnet, and in places biotite, albite, and ankerite.

TOURMALINE: In two small areas in the no. 9 ledge structure, tourmaline occurs with quartz of the first stage of mineralization. There is also a considerable amount of pyrrhotite, interstitial and veining the tourmaline.

BIOTITE: Coarse black hydrothermal biotite accompanies small quartz veins in cummingtonite schist, and green hydrothermal biotite is widespread throughout the mineralized portions of the Homestake formation. Dodge (1935; 1942) determined that the optical properties of both varieties were within the range of the biotites in the unmineralized schists. The coarse black biotite analyzed by Sharwood (1911, p. 752, 753) was undoubtedly hydrothermal. The biotite replaces the schist minerals cummingtonite, quartz, sideroplesite and garnet and in a few places replaces vein quartz, chlorite, albite, and perhaps arsenopyrite.

GARNET: Dark-red hydrothermal garnet replaces the schist minerals cummingtonite and quartz and the hydrothermal minerals quartz and chlorite. Dodge (1935; 1942) determined the indices to be within the range of the almandite in the schists, and Sharwood (1911, p. 753) analyzed a hydrothermal garnet which was essentially almandite.

CUMMINGTONITE: Hydrothermal cummingtonite replaces hydrothermal quartz, garnet, and albite. This cummingtonite is lighter in color than the cummingtonite of the regional metamorphism, some being almost white; but the lighter color probably is due to a greater degree of twinning, because the optical indices are within the range of the indices of the regional cummingtonite.

ALBITE: Pale-green albite replaces hydrothermal quartz, chlorite, biotite, garnet, and cummingtonite. In a few sections it is replaced by chlorite, biotite, and cummingtonite, which probably indicates a repetition or overlap.

ANKERITE: Ankerite replaces the schist minerals cummingtonite, biotite, quartz, and garnet, and the hydrothermal minerals quartz and chlorite. A very little chlorite is later than ankerite. Some minute veins of carbonate which cut arsenopyrite have both ankerite and calcite, but most of these veins have only calcite.

SERICITE: Sericite, with chlorite and calcite, replaces the garnet of the schists and hydrothermal quartz, chlorite, ankerite, and albite. Tiny plates of sericite fill a few fractures in arsenopyrite, pyrrhotite, and pyrite.

CALCITE: Five forms of hydrothermal carbonate have been identified: ankerite, rhodochrosite, dolomite, iron-bearing calcite, and (most abundant) pure calcite. Ankerite is locally abundant in some parts of the ore zone. Dolomite is found in many quartz veins in the Poorman formation. The iron-bearing calcite, in which ω ranges from 1.661 to 1.668, occurs with pure calcite in a few of the calcite-pyrite veins. Rhodochrosite has been found in only three or four places in calcite-pyrite veins. Carbonates in specimens of hydrothermal replacements of the Homestake formation can usually be assumed to be either calcite or ankerite, but both are found in many specimens. Minute grains of carbonate in thin section cannot, therefore, be determined with certainty. Without doubt, however, calcite replaces cummingtonite and biotite of the schists and hydrothermal quartz, chlorite, and albite. Tiny veinlets of calcite cut arsenopyrite and pyrrhotite quite abundantly and cut pyrite in a few instances.

Summary of the Stages and Sequence of Mineralization

The probable qualitative relations between the changes in pressure on the liquids occupying the open spaces in the rocks and the changes in composition of the mineralizing solutions entering the Homestake formation are shown in Figure 4. Increase in pressure was caused by the stresses that fractured the rocks. Since

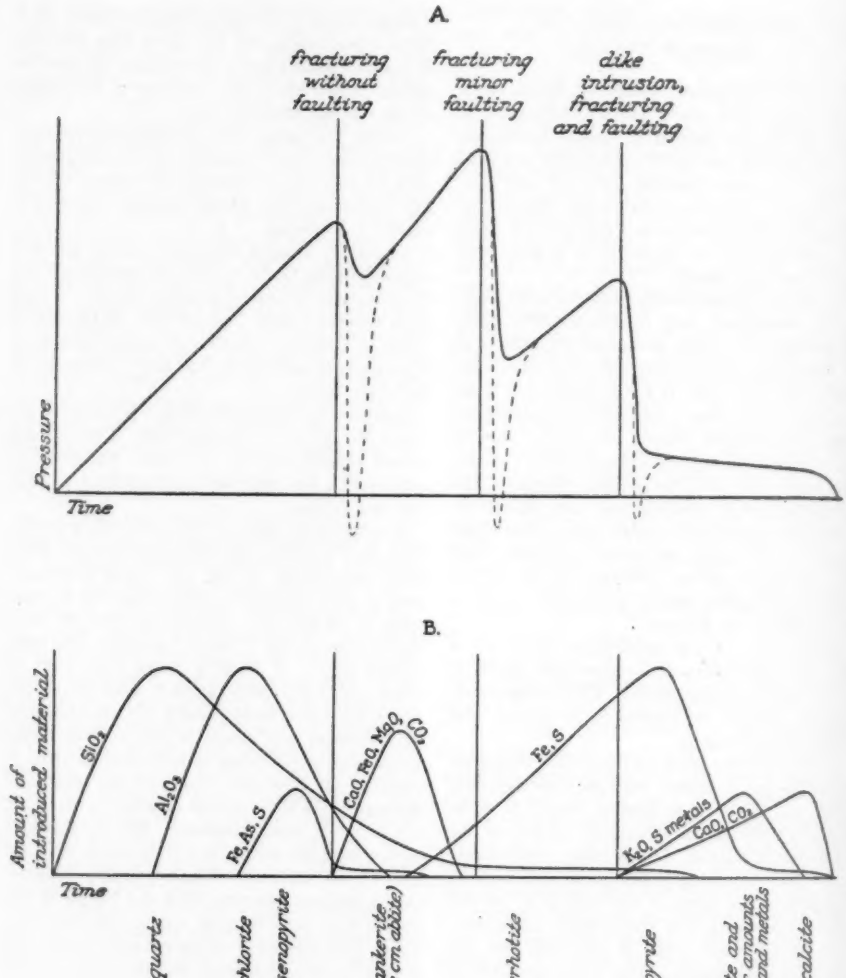

FIGURE 4.—CHANGES IN PRESSURE AND COMPOSITION OF MINERALIZING SOLUTIONS
A. Changes in pressure on solutions in openings in the rocks
B. Changes in material introduced into the Homestake formation

flow in the open spaces in the rocks was hindered, the pressures on the liquids locally were higher than the average pressures, and after fracturing the pressures on the liquids were appreciably less near the fractures than elsewhere and probably were less than the original pressure.

The substances deposited in great amount in the Homestake formation were, chronologically, silica, alumina, lime, iron, sulphur, more lime, and carbon dioxide. In addition very large amounts of volatile constituents and soluble salts must have passed through the formation without leaving any record. Most

of the large amount of vein quartz in the mineralized portion of the Homestake formation was introduced, although some of it was merely recrystallized from pre-existing quartz or silicate minerals. The extensive replacement of cummingtonite by chlorite (daphnite) can have resulted only from a large addition of alumina to the Homestake formation and a large removal or redeposition of silica. Arsenopyrite formed by the introduction of a little arsenic, sulphur, and perhaps iron. Introduction of lime and probably of iron, magnesia, and carbon dioxide formed ankerite. The constituents for the small amount of hydrothermal biotite, garnet, and cummingtonite were probably derived from the same minerals in the rocks, but albite was a result of a small addition of soda. A large amount of iron and sulphur went into the formation of pyrrhotite and pyrite. Specularite and magnetite may have distilled from the magma of the rhyolite dikes during the widespread fracturing of the rocks at the time of the dike injection (Burbank, 1936; Wells, 1938; Loughlin, 1941). Although no overlap of specularite and magnetite by pyrite has been detected, possibly the deposition of pyrite was merely interrupted by a short period of deposition of iron oxides. When falling temperatures temporarily permitted the formation of small amounts of acid solutions (Bowen, 1933, p. 106–128), some marcasite was formed by alteration of earlier sulphides. The final stage of mineralization closed with the introduction of a large amount of lime and carbon dioxide and minor amounts of potash, fluorine, antimony, sulphur, copper, iron, zinc, lead, and gold. The deposition of gypsum near the end of this stage indicates that the temperature of the solutions was somewhat less than 66° C. (Langford and Hancox, 1936, p. 604, 605), and not much above the present temperature of the rocks of the lower mine levels. It also indicates that there was an abundance of SO_4 radical, an inference which is probably verified by the analysis of water from a crevice. The close association of sulphates and hydrogen sulphide in the waters of these crevices probably supports the suggestion of Wandke (1925) and Graton (1933; Graton and Bowditch, 1936) for formation of sulphuric acid by reaction of H_2O and S; but the reaction took place here at the very end of the hydrothermal process, probably at low temperatures, and it is doubtful whether any free acid formed in the presence of considerable amounts of alkaline carbonates.

Ross (1935) has shown that in the pyrrhotite veins of the southern Appalachians a number of hydrothermal stages are separated by structural breaks and changes in composition of the ore solutions. The succession in those veins is similar to that in the Homestake deposit; the main differences are that in the southern Appalachian veins a single stage of carbonates intervenes between two stages of silicate minerals, pyrite is the earliest sulphide whereas arsenopyrite is the latest, and chlorite is later than the other silicates and perhaps later than the sulphides. Park and Wilson (1936) have described a hydrothermal recrystallization of metamorphic minerals somewhat like the silicate stage of the Homestake mineralization.

Unsolved Problems

Age of the Gold Mineralization

Although it is scarcely possible to describe the Homestake gold deposit without becoming involved in a discussion of the age of the ores, it is not the purpose of this paper to give a final answer to that question, and the age of the Homestake gold mineralization rightly heads the list of the unsolved problems. There are valid arguments both for pre-Cambrian and for Tertiary age of the ores, but the answer is probably not that there are two ages of mineralization, for there are also valid arguments for there being only one main period of mineralization. Obviously then some of these arguments are wrong; but neither the pre-Cambrian nor the Tertiary arguments are yet strong enough to silence the opposing view.

Evidence supporting a pre-Cambrian age are: (1) There is detrital gold in the basal Cambrian conglomerate adjoining the Homestake ore bodies; and (2) The Homestake ores are of so much more deep-seated character than the known Tertiary ores of the adjacent Bald Mountain district that they must have been formed much earlier under much deeper cover. Evidence supporting a Tertiary age are: (1) The ores have been deposited in a

sequence of interrelated stages, and the latest stage is later than the Tertiary dikes; (2) The areal distribution of the Homestake deposit and the Tertiary Bald Mountain deposits is too nearly identical to be accidental; and (3) The cross folds in which the Homestake ores are localized were caused by the intrusion of the Tertiary igneous rocks. Each line of evidence is reviewed in some detail, if for no other reason than to call attention to the difficulty involved in determining the age of any ore deposit in pre-Cambrian rocks.

The concentration of gold in the basal portion of the Cambrian Deadwood formation attracted early attention and was quickly used as proof that the Homestake ores were pre-Cambrian (Devereux, 1882; Irving, Emmons, and Jaggar, 1904). The weight of evidence from these descriptions leaves little doubt that these are "fossil placers", although Irving (Irving, Emmons, and Jaggar, 1904, p. 98–111) did note that some of the ore was a pyritic replacement of the conglomerate. Although all work in these ores ceased many years ago, the workings are still open except at the portals of the tunnels, and all the workings have been opened and restudied. This study casts so much doubt on the placer origin of the gold that this line of evidence can have little weight. The mineable portions of the deposits are the oxidized portions of leaner pyritic replacements of the conglomerate, and the space distribution of the gold which originally suggested a placer origin is probably attributable to a small amount of solution and redeposition of gold during oxidation. These pyritic replacements are identical with the pyritic replacements of Cambrian dolomite in the Tertiary ores of the Bald Mountain district, a few miles to the west.

Evidence supporting a pre-Cambrian age of the Homestake ores because of their deep-seated or hypothermal character has more to commend it. The Homestake ores will certainly be classed as hypothermal by anyone who uses the depth-zone classification, whereas the Tertiary ores of the Bald Mountain district would be classed less certainly as mesothermal, but certainly of lower intensity than the Homestake ores. This line of evidence suggests that the two types of ore formed at different depths, hence at different times; therefore, one may be pre-Cambrian and the other Tertiary. The reasoning is valid if the depth-zone classification is valid. A point in its favor is the position of pyrite in the sequence of minerals, not at the beginning of the sequence as it normally occurs but definitely later than arsenopyrite and pyrrhotite. A point unfavorable to this line of evidence is that specularite and magnetite, which normally are considered hypothermal or pyrometasomatic, are definitely post-rhyolite and so cannot be included as part of the supposed pre-Cambrian hypothermal mineralization.

The most compelling evidence supporting a Tertiary age of the Homestake ores is that the mineralization was effected in a series of separate but interrelated stages, and the last stage was later than the Tertiary rhyolite dikes. The last stage is certainly later than the dikes, but the interrelation of the four stages into a single course of mineralization is questionable. Evidence for interrelation of the separate stages is: (1) transitions by overlap or repetition of minerals; (2) separate stages in identical fracture patterns, and the same areal distribution for separate stages; and (3) vertical zoning whereby pyrrhotite disappears in the upper levels, its place being taken by pyrite. A point which weakens the argument is that the writer has concluded that gold is late forming, as late as pyrite and calcite or later, but it shows no structural control by the late pyrite-calcite veins (only minute amounts have been found in pyrite-calcite veins). The fact that gold though late forming shows a strong association with the early-forming arsenopyrite is also a point of weakness unless it can be explained by a chemical control. If the description of the four stages of mineralization is accurate but the interrelations are denied, then the first three stages of mineralization might be called pre-Cambrian, the fourth one Tertiary. Where the gold belonged would depend on whether one or several periods of introduction of gold were assumed; at least a small amount of gold is with the fourth stage and so has to be Tertiary.

The Tertiary gold ores of the Bald Mountain district are replacements of Paleozoic rocks, principally the Cambrian Deadwood formation. They occur in an annular ring almost

completely surrounding the pre-Cambrian core of the small dome of the Deadwood-Trojan uplift (Noble, Harder, and Slaughter, 1949). The Homestake ore bodies are in pre-Cambrian rocks but lie in this same annular ring. Probably a Tertiary mineralization would not be so accurately superimposed on a pre-Cambrian mineralization.

Another line of evidence, which may become conclusive but which cannot be so classed now, tries to relate the ore-bearing cross folds to deformations caused by the intrusion of Tertiary igneous rocks. The cross folds have the direction of movement that might be expected from the intrusion of the igneous rocks, and there is no evidence that they could not have been produced by the intrusions. There is abundant evidence that the dome of the Deadwood-Trojan uplift was produced by Tertiary intrusive activity (Noble, Harder, and Slaughter, 1949). The direction of displacement in the cross folds of the main ore body, no. 5 ledge ore body, and the ore bodies of the no. 9 ledge is west side up or south, such as would be produced by the intrusion of the Cutting stock, northwest of the mine area. The direction of displacement of the cross folding in the Caledonia ore body is east side up or south, presumably related to another stock to the northeast. The amount of displacement of the pre-Cambrian rocks in the zones of cross folding seems to be much greater than any corresponding displacements in the overlying Cambrian strata, but a study of the smaller cross folds shows that the cross-folded zones are discontinuous, and the displacements quickly die out at the ends of the zones. Moreover, small monoclinal flexures in the Cambrian strata are in the same direction as the displacements in the pre-Cambrian rocks. The answer to the question whether the cross folds are pre-Cambrian or Tertiary may come when the ore-bearing folds of the no. 9 ledge structure are followed to the base of the Cambrian strata.

Difficult as it is to fix time relations in this deposit and in others like it, the significance of the time relations may be misinterpreted. The question of the age of the Homestake mineralization in relation to the Tertiary rhyolite dikes has been a fruitful source of argument, and it seems to have been implied that the answer to this question would fix the age of the ores. If all the mineralization were post-dike, which is not true, then the ores would be Tertiary; but when it is proved that part is pre-dike and part is post-dike, nothing has been proved, because the pre-dike portion can be either pre-Cambrian or Tertiary. The answer to the question whether the cross folds are pre-Cambrian or Tertiary likewise may fail to determine the age of the ores. If the cross folds are shown to be Tertiary, then the ores are Tertiary, since they are localized by the cross folds. But if the cross folds are pre-Cambrian, the ores can then be either pre-Cambrian or Tertiary.

Localisation of Gold

An unsolved problem for which we have no immediate answer is the question of why the gold is localized in the Homestake formation. The Poorman formation is dominantly the carbonate ankerite or ankeritic dolomite, with some bands of recrystallized chert and a little argillaceous matter. The Homestake formation, overlying the Poorman, was originally the iron-magnesium carbonate sideroplesite, but by regional metamorphism the carbonate was converted in part of the district to the iron-magnesium silicate cummingtonite. There are also bands of recrystallized chert and a very little argillaceous matter. The Ellison formation overlying the Homestake is dominantly argillaceous phyllite with considerable quartzite. In all the formations there is a considerable amount of finely divided graphite. The gold is almost wholly localized in the Homestake formation. A very small amount, not sufficient for a separate mining operation, occurs in the Poorman formation, but practically no gold is found in the Ellison formation. The localization of gold in the Homestake formation can scarcely be structural, related to the strength of the rocks or the nature of the fracturing, because the gold replaces both the carbonate sideroplesite and the silicate cummingtonite, and these minerals are as distinct in physical properties as any of the minerals in the other formations. If the localization is not structural, it is probably chemical, but the chemistry is not easily ex-

plained. About the only chemical feature common to the sideroplesite schist and cummingtonite schist is the iron-magnesium ratio in the two minerals. Why the underlying Poorman formation, with an abundance of iron, magnesium, and lime, was not mineralized in important amounts at the time is an unanswered question.

A related question is why gold is localized in the vicinity of arsenopyrite. Microscopic study indicates a considerable time interval between the early arsenopyrite and the late gold; yet arsenopyrite is the best visual guide to ore in the mine, and the microscope shows that specks of gold occur most frequently in the near vicinity of arsenopyrite. Schwartz (1944) noted a similar control. Here again the control is probably chemical, but the chemistry of the process is unexplained.

Dilatation

The adequacy of the principle of dilatancy to explain the localization of Homestake ore bodies is questioned because the concept needs further study. The principle appears to be valid, but whether the amount of permeability developed by dilatation is adequate to control the movement and deposition of ore-bearing solutions is not established.

Stages of Mineralization

A fundamental question, not only for the Homestake ore bodies but for many other ore bodies, is the reason for the deposition of the ore-forming constituents in a series of stages. Why is there a "normal sequence" of mineralization? In one ore deposit after another ore minerals were deposited in a sequence which is nearly the reverse of the order of deposition expected if supersaturation of sulphides were the control. In general, the most soluble sulphides form first, the least soluble last. Tendency for ore deposits to show a normal mineral sequence not in the order of insolubility must mean that the controlling factor is a change of composition of the ore-bearing fluids at their source or gathering place.

Changes of temperature are generally considered much more important than changes of pressure in controlling the deposition of minerals from hydrothermal solutions. Theoretically, and over the ranges of temperature and pressure normally used in experimental work, this is probably true. However, pressures on the hydrothermal solutions when magmas are deforming the rocks by forcible intrusion may be vastly greater than those referred to in comparing the effects of temperature and pressure on hydrothermal solutions. In the vicinity of the Homestake mine, the intrusion of Tertiary igneous rocks raised the superjacent cover of 8000 feet of Paleozoic and Mesozoic rocks several thousand feet. Any hydrothermal solutions circulating in the rocks before the rocks became thoroughly fractured would be subjected to differential stresses measured not by the weight of the 8000 feet of rocks, but by their resistance to deformation. Step by step failure of the rocks would then be matched by corresponding changes in the pressures on the hydrothermal solutions. In some way the effect of these changes seems to operate at the source or gathering place to control the composition of the hydrothermal fluids at that place.

Changes of pressure on the hydrothermal solutions may explain some of the puzzling relations of pyrrhotite and pyrite in the Homestake ores. In most of the Homestake mine the pre-dike iron sulphide is pyrrhotite, the post-dike is pyrite. Still later there is a little marcasite. The time distinction between pyrrhotite and pyrite is sharp enough to permit belief, by one line of reasoning, that the earlier sulphide is pre-Cambrian, the later one Tertiary. Certain relations, however, suggest that factors other than a wide difference in age may be involved. In several places, narrow pyrrhotite veins of the third stage of mineralization cross from Homestake formation into Ellison formation, and in each place the pyrrhotite vein within the distance of a few feet changes to a pyrite vein; the pyrrhotite disappears. A similar change has been noted in a few places where the pyrrhotite veins cross masses of vein quartz. Within the quartz itself the sulphide is pyrite, but beyond the quartz it is pyrrhotite. On a broader scale the same relation prevails, because pyrrhotite is the most abundant sulphide in both the Homestake and the Poorman formation but is virtually ab-

sent from the Ellison formation, although pyrite is fairly abundant in places in the Ellison. A fundamental distinction between the Ellison formation and the other two formations is that the Ellison has much more quartz. The control between pyrrhotite and pyrite may be chemical, but it may also be structural, if the quartz-rich Ellison formation was more thoroughly fractured when the ore solutions entered.

That the control between pyrrhotite and pyrite was structural rather than chemical is suggested by the vertical distribution of the minerals. About on the 800 level in the main ore body and about on the 400 level in the Caledonia ore body pyrrhotite disappears, and its place is taken by pyrite. In this part of the mine area the base of the Cambrian Deadwood formation is some 200 feet above the zero reference level; hence the disappearance of pyrrhotite takes place in the pre-Cambrian rocks some 600 to 1000 feet below the base of the Cambrian rocks. This is many hundreds of feet below any effects of Cambrian weathering and is wholly unrelated to any change in composition of the rocks. It may be related, however, to permeability and pressure, since the overlying Paleozoic rocks can be assumed to have been much more permeable than the pre-Cambrian schistose rocks. The upper few hundred feet of the pre-Cambrian rocks may have contained enough leaks to permit the tremendous pressures developed in depth by magma intrusion to fall off near the more permeable rocks.

A study of the sequence of mineralization in the Homestake mine furnishes further evidence for the conclusion stated elsewhere (Hulin, 1930; Hewett, 1931; Ferguson and Gannett, 1932; Park and Wilson, 1936; Lindgren, 1937; Mawdsley, 1938; Galbraith, 1941; White, 1943) that gold is usually the last or nearly the last mineral in the sequence of ore deposition. The reason must be not that gold is the most soluble constituent in the hydrothermal solutions, but that it is the last mineral to collect at the source or gathering place.

The sequence of stages of mineralization in the Homestake ores raises some question regarding the validity of the depth-zone classification in common use in the study of ore deposits. By the commonly accepted criteria of mineral-stability ranges, and assuming the four Homestake stages are parts of one continuous process, the Homestake mineralization went from hypothermal through a brief pyrometasomatic stage (the biotite-garnet phase of the second stage), back to hypothermal (to the end of the third stage), then through a brief pyrometasomatic stage (magnetite and specularite) to mesothermal, and finally to epithermal (marcasite and gypsum). Gold can be interpreted as belonging to all stages, but the writer believes the gold came in only during the final or epithermal stage. How, then, shall we characterize the Homestake mineralization? On the basis of the highest intensity reached, it is pyrometasomatic. On the basis of the most abundant sulphides (pyrrhotite, pyrite, and arsenopyrite) it is hypothermal. As a gold deposit, however, it is epithermal, if gold formed only in the epithermal stage. The same question has been raised in many other places (Buddington, 1930; 1935; Loughlin and Behre, 1933; Lindgren, 1937; Galbraith, 1941; Shafer, 1945), because this descending intensity range is probably the rule rather than the exception. To the writer the answer, for gold deposits and perhaps for all ore deposits, lies in a change from the depth-zone classification to something like the classification of De Launay (1913, p. 1–54), based on the composition of the ore-bearing fluids.

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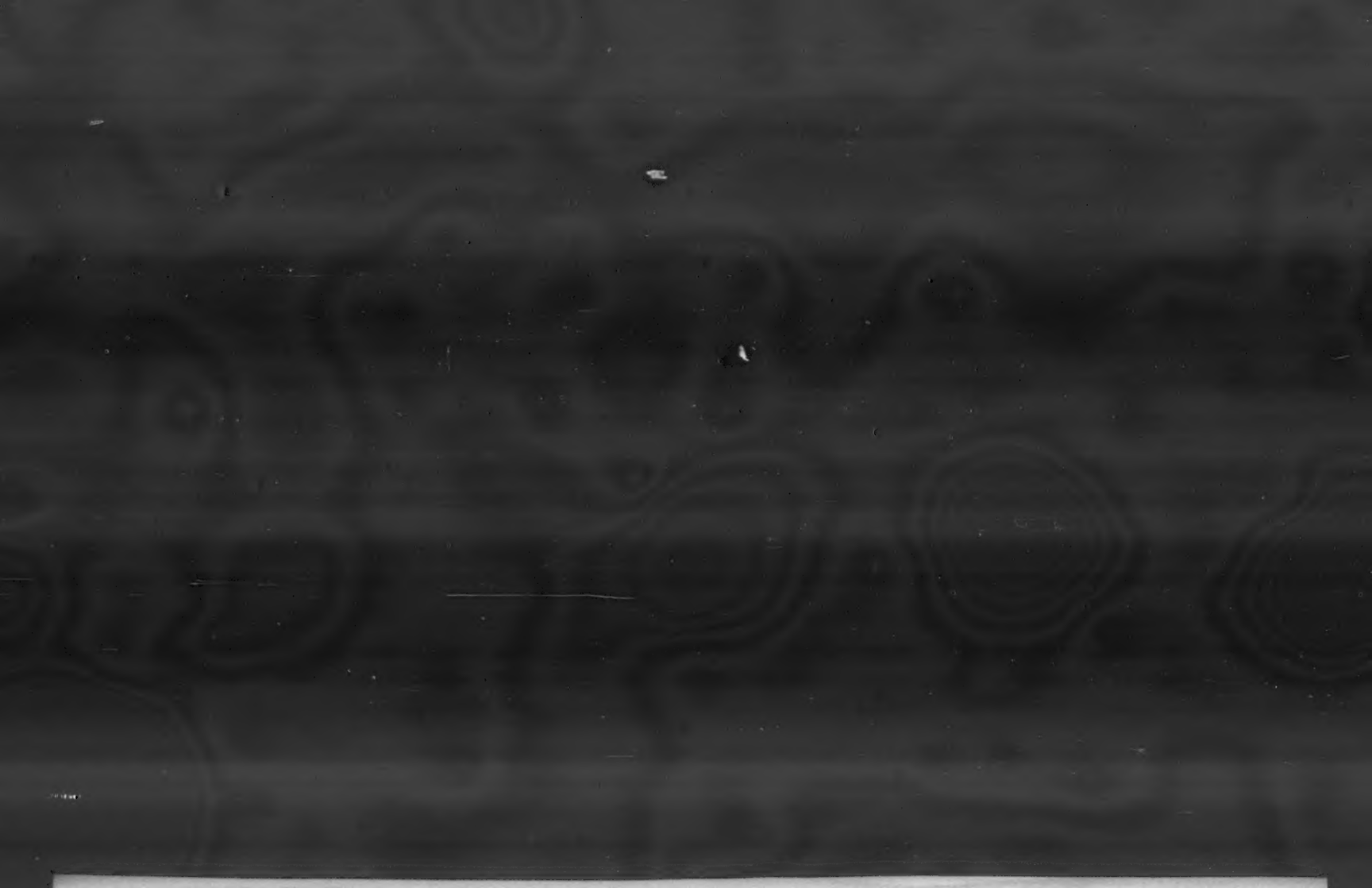

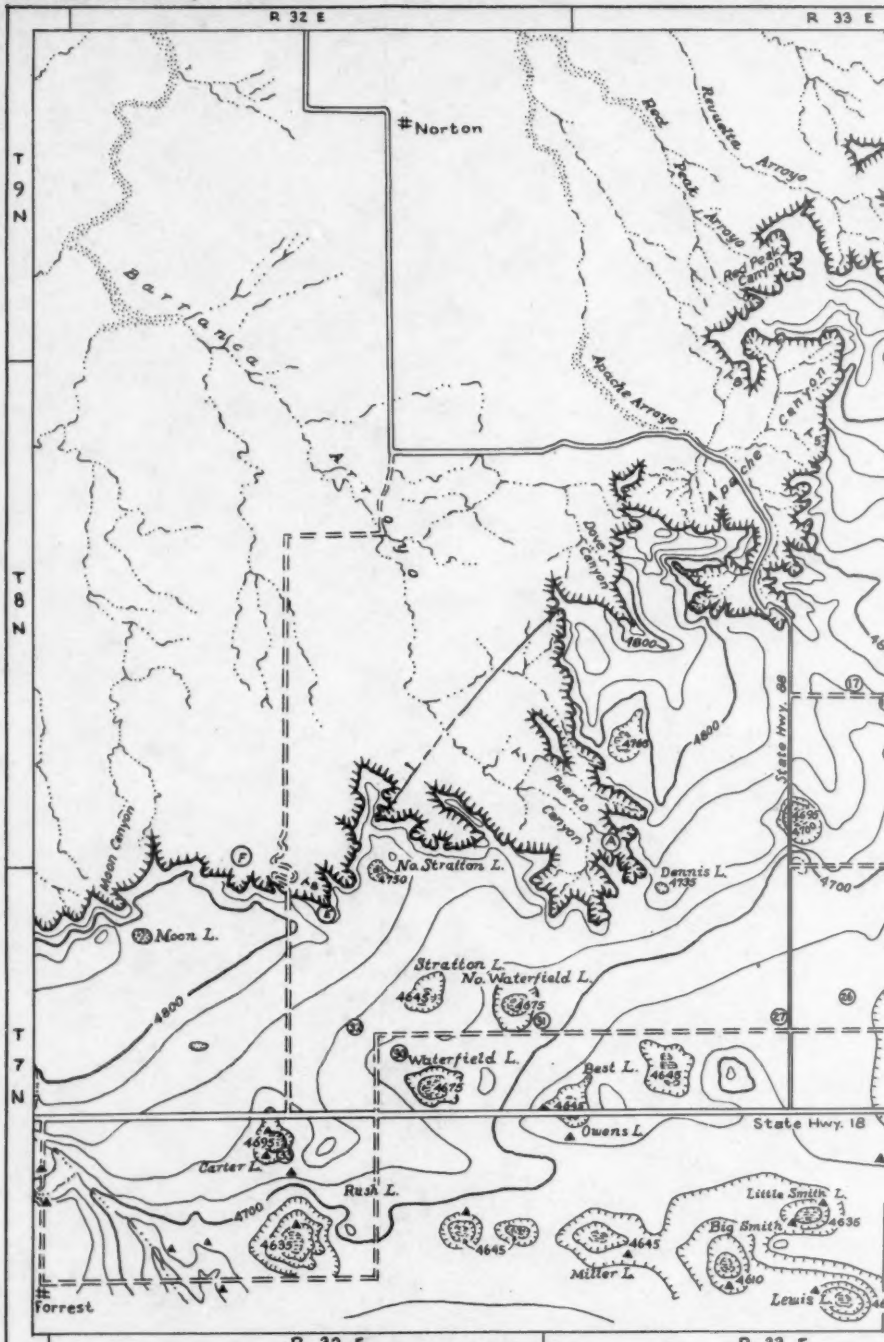

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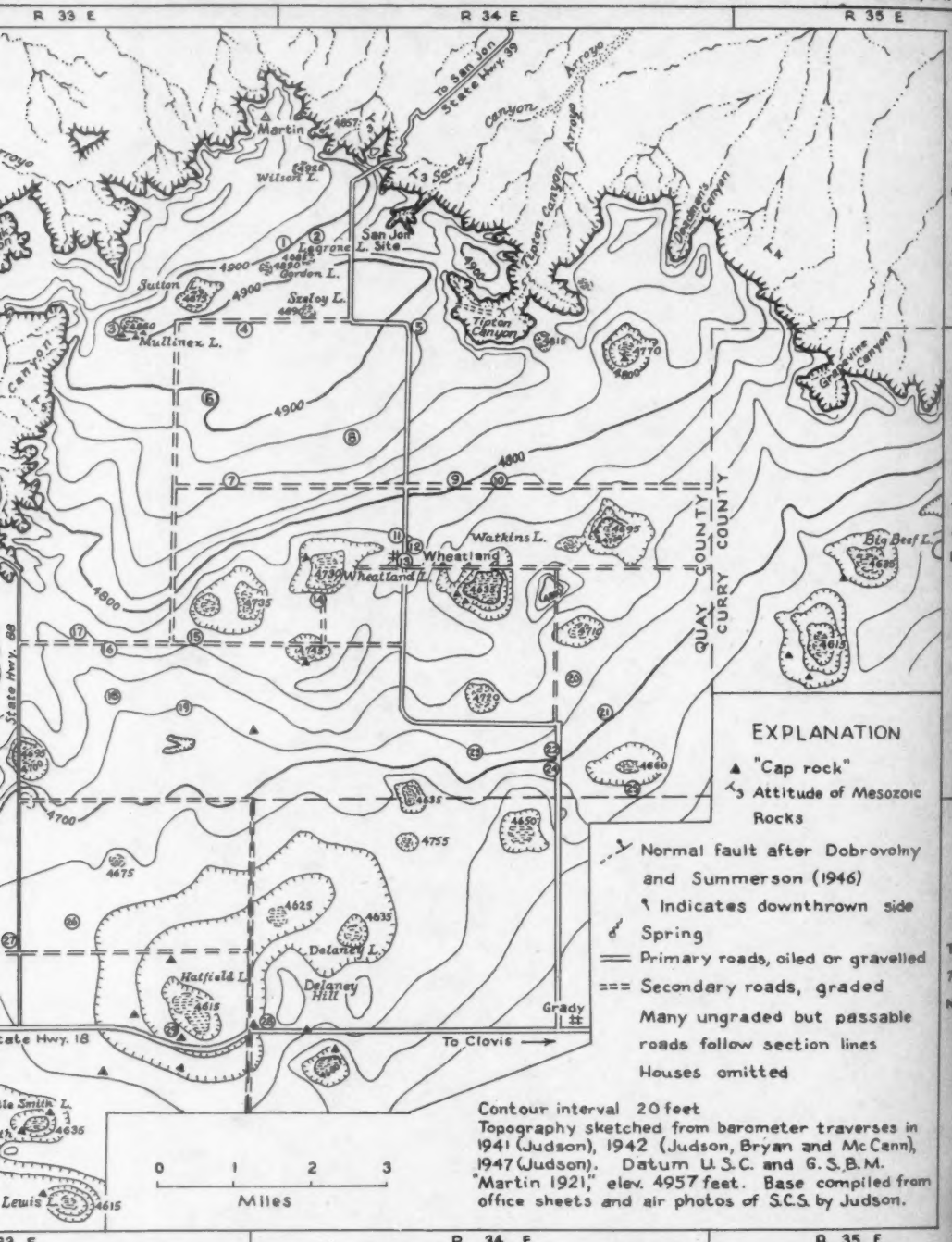

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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DEPRESSIONS OF THE NORTHERN PORTION OF THE SOUTHERN HIGH PLAINS OF EASTERN NEW MEXICO

By Sheldon Judson

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Abstract

Study of the shallow depressions of the Southern High Plains of eastern New Mexico shows that they are the result of alternate periods of leaching and wind deflation and not of collapse into the underground or local subsidence in the Pliocene sediments. During wet periods of the Pleistocene the calcareous cement of the Ogallala formation was destroyed. during succeeding dry periods these locally leached areas suffered wind deflation. The depressions are located along broad shallow troughs which involve the "cap rock" of the Plains in a manner not yet clearly understood. Evidence suggests that the "Pliocene cap rock" may not have the value as a stratigraphic horizon that is usually attributed to it. The Cuneva depression, a true collapse depression active within historic times, is described.

Introduction

General Statement

The High Plains, stretching from Texas and New Mexico on the south to the Pine Ridge Escarpment in Nebraska and South Dakota on the north, represent large, discontinuous remnants of a once vast and unbroken apron of stream-laid sediments derived from the rising

Rocky Mountains in Tertiary time. The upper surface of this apron was graded to the Late Pliocene seas of Texas and to the major drainages of the interior continental basin. The Northern, Central, and Southern High Plains, residuals from this sedimentary apron, have surfaces which approximate those across which coursed the Late Pliocene streams. Broad regional uplift and gentle deformation have somewhat distorted the original plain. Furthermore, mantles of eolian material in many places raise its level, and countless undrained depressions, containing ephemeral wet-weather lakes, lie below its original grade.

This study of the depressions of the High Plains was prompted by the investigation of the geologic antiquity of the San Jon site, 10 miles south of the town of San Jon, New Mexico (Roberts, 1942). The site is located in a dissected depression along the northern edge of the Southern High Plains. Information concerning the genesis of this and neighboring depressions, both dissected and undissected, is a by-product of the original investigation.

Acknowledgments

The writer is indebted to the authorities of the Smithsonian Institution who made available funds for the summer field season of 1941 and 1947 as well as for a two-week survey in 1942. The continuing interest, co-operation and assistance of Dr. Frank H. H. Roberts, Jr., Smithsonian Institution, has greatly facilitated the geological work in both field and office. Prof. Kirk Bryan, Harvard University, supervised the field work and criticized this paper in manuscript. Dr. Franklin T. McCann, Auburn University, accompanied Bryan and Judson on the 1942 trip. The writer's wife, Anne Perrin Judson, served as field assistant in 1947. Prof. Lewis M. Cline, University of Wisconsin, has kindly criticized the manuscript, as has Dr. W. Armstrong Price.

Historical Summary

Five major theories have been advanced to account for the origin of the undrained depressions of the High Plains: (1) deep-seated solution followed by complete or partial collapse of the surface (Johnson, 1901; Elias, 1931; Theis, 1932; Smith, 1940; Frye and Schoff, 1942); (2) differential subsidence within the Pliocene (Johnson, 1901; Schoff, 1939; Frye and Schoff, 1942; Frye, 1945a); (3) deflation (Gilbert, 1895; Evans and Meade, 1945); (4) "Caliche karst" (Price, 1940); and (5) removal of material by animals (Passarge, 1904; Darton *et al.*, 1915). Only the first four are receiving the attention of current workers. The present study had, by 1942, led to the same general conclusions presented by Evans and Meade (1945). A detailed mechanism and a considerable body of facts amplify the generalities presented by them.

Topography of the Plains

The High Plains slope eastward in the general direction pursued by the streams of Tertiary time. The slope is presumably steeper than the original gradient of Tertiary streams by reason of uplift. In the Southern High Plains the slope is to the east-southeast. Several streams such as the Tierra Blanca and the Blanco have cut valleys through the Tertiary into the underlying rocks. In addition the long depression known as the Portales Valley extends southeasterly from the western border of the Plains. The general slope of the plains, however, as well as the valleys of major streams and the greater depressions like the Portales Valley, require for their interpretation the study of a larger area than here attempted.

The minor irregularities of the Plains, the depressions, vary in size, shape, and depth. Some are but a few yards wide and less than a foot deep. Others are more than a mile in diameter and over 50 feet deep. These depressions contain ephemeral lakes, some of which are perennial during a series of wet years and dry up only after a series of dry years.

The Northern Escarpment of the Plains facing the Canadian Valley offers advantages to the study of the depressions. The San Jon site is an obvious depression breached by the relatively recent recession of the Escarpment. The pattern of the edge of the "cap rock" (Pl. 1)[1]

[1] The roads and section lines of the map are copied from maps of the Soil Conservation Service. The edge of the "cap rock" of the Plains is compiled from air photographs of the same organization. The contours were sketched in the field at a scale of 1 inch to the mile and based on elevations secured by

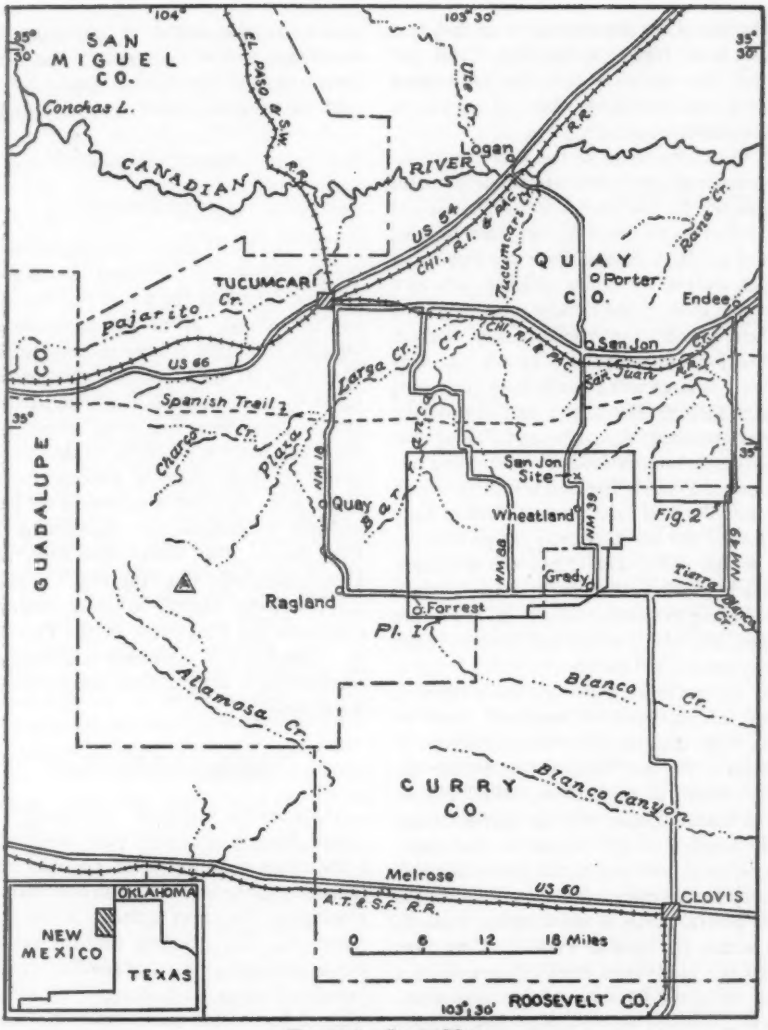

FIGURE 1.—INDEX MAP
A indicates location of the Cuneva collapse depression.

indicates other breached depressions. Thus Tipton Canyon has a bowl shape that opens to the plains below through a narrow gap. Other canyons that indent the border of the plains have a similar form and origin. The area in the vicinity of the San Jon site thus affords an opportunity to study not only the shape and

traverse with a Paulin altimeter in 1941, 1942, and 1947. Elevations are referred to a Coast and Geodetic Survey triangulation station and benchmark, "Martin, 1921", elevation, 4957 feet. Individual points were determined within 10 feet. Because the traverses were made mostly on roads following section and half-section lines, the determined points are sparse, and the contours lack precision. It is believed, however, that they show fairly well the general shape of the land and some of the significant detail. Near the escarpment the topography is quite complex, and contours are omitted or have only sketch value.

distribution of the depressions in an area that appears to be typical of the High Plains, but also provides exposures along the Escarpment in which the internal structure of depressions may be studied.

On the Plains south of the Escarpment depressions trend east-northeast, parallel to major troughs (Pl. 1). The plain slopes southeast and thus conforms to the well-recognized regional slope of the High Plains. There are four major troughs marked by chains of lakes each in a minor and more or less circular depression. The northern trough, the smallest, runs through the San Jon site to Apache Canyon. The next southerly is large and extends from Grapevine Canyon through Wheatland and toward the southern portions of Apache and Puerto canyons. It may be conceived of as extending farther toward Forrest. The next southerly trough runs from Big Beef Lake through Hatfield Lake and toward the northeast may merge with the next trough north. The fourth line of depressions is barely indicated on the map. Grady is located along its northern slope, but it was not mapped, and its full extent and relation to the other troughs is not known.

The east-northeast trend of lakes is replaced by an east-southeast or southeast trend at places. Field inspection and a consideration of the maps of the Soil Conservation Service (U. S. Department of Agriculture, 1936–1937) indicated chains of lakes with an east-southeast trend, parallel to the course of the major streams, south and east of the area studied. A tendency for drainage to develop in lines normal to the general strike of the contours is shown even within the area of Plate 1. Thus, near Forrest the headwaters of the Blanco lie in a valley cut in the Pliocene and trending southeastward. A valley with a similar trend extends from the head of Apache Canyon into the large depression containing Hatfield Lake. A smaller but more sharply defined drainage enters Big Beef Lake from the north-northwest.

These troughs and the broad ridges described above appear to involve the "cap rock". It crops out here and there on both sides of the troughs (Pl. 1). As shown in logs of wells and by observations along the rim of the Plains, the "cover" over the "cap rock" ranges in thickness from a thin film to 40 feet. The broad ridges and intervening troughs cannot be accounted for by variations in the thickness of the "cover". Broad undulations in the "cap rock" must exist—a situation discussed more fully below.

The "Cap Rock"[2]

General Statement

The "cap rock" crops out continuously along the Escarpment of the Plains. To its position as a cliff-former at the top of this Escarpment and immediately above the sandstone of the Ogallala formation, it owes its designation as a "cap rock". It underlies the Plains almost continuously and is exposed at points, particularly around depressions and along stream valleys. This "cap rock" is generally considered to represent the final phase of Pliocene deposition and to mark the boundary between the Pliocene and the Pleistocene. (*Cf.* particularly Lugn, 1939; Smith, 1940; Elias, 1942; Frye, 1945b.) Physiographically this "cap rock" is, perhaps, not incorrectly named. Whether, however, it represents the final event of the Pliocene, or even whether it is a continuous lithologic unit representing a definite time interval are questioned below.

Lithology and Distribution

At its top the "cap rock" is characterized by a hard, dense, platy almost pure limestone from a few inches to 10 feet thick. On fresh surfaces it is flesh pink to buff or whitish and weathers to a dull gray. The upper surface is in many places botryoidal. The limestone contains concentric banding, concretionary and pisolitic structures, brecciated zones, and stringers of secondary silica. A few frosted quartz grains are scattered through it. Locally bioherms up to 6 or 8 feet

[2] Lugn (1939) has raised the Ogallala to the rank of group and included within it four separate formations. The youngest, the Kimball formation, includes the "cap rock" and the sandstone immediately below it. Elias (1942) follows this classification and refers to the upper platy limestone of the "cap rock" as the "algal limestone". Frye (1945b) also employs "algal limestone" for this zone of the "cap rock". Despite its shortcomings the writer retains in this discussion the long-used term, "cap rock". Its use, though clumsy at times, is not impossiple and pending a clearer understanding of the "cap rock" it is deemed the best available term.

in maximum dimension are present. In many other places the limestone shows signs of organic deposition as reported in Kansas and Nebraska by Elias (1931) and from the High Plains of Texas and New Mexico by Price, Elias, and Frye (1946). The platy, upper zone of the "cap rock" overlies a sandy granular limestone grading downward into concretionary and nodular sandy limestone and sand with concretions. This lower portion is essentially a white to buff caliche and ranges in thickness from a few inches to 10 feet. It grades downward into the laminated and bedded sandstones of the Ogallala formation. These sandstones have a crystalline, calcareous cement and massive concretions ranging up to 2 feet in diameter.

The "cap rock" with its platy upper zone of almost pure fresh-water limestone is usually considered the true "Pliocene cap rock" of the Ogallala formation. There are, however, plates of limestone or other "cap rock". Some are considerably younger than, and lithologically different from, the "Pliocene cap rock". Others, however, are indistinguishable from it.

"Cap rocks" obviously younger than the "Pliocene cap rock" and of undoubted Pleistocene age occur along the Escarpment and are particularly well developed in Puerto and Deadman's canyons. In one of the prongs of Puerto Canyon (*A*, Pl. 1), a Pleistocene "cap rock" makes an inner cliff and occurs on several projecting points (Pl. 2, fig. 1). At its lowest altitudes it consists of a platy limestone about 3 feet thick. The rock is gray and lacks the pink color and brecciation characteristic of the algal limestone forming the upper zone of the "Pliocene cap rock." It grades downward into massive caliche, and the two lie on about 10 feet of unstratified gravel. The pebbles are mostly rounded fragments of the "Pliocene cap rock", but there are also pebbles and fragments of pebbles derived from the gravel beds of the Ogallala. Several of these pebbles were observed to have finely pitted, fluted and polished surfaces. They are ventifacts. The elevation of this Pleistocene "cap rock" increases laterally in all directions toward the outcrops of the "Pliocene cap rock". It was obviously deposited in a depression below the general level of the Plains. Because the upper zone of the platy Pleistocene limestone appears to be confined to the more interior outcrops and at lower elevations it presumably marks the area most commonly covered by lake waters.

At the head of Deadman's Canyon are additional exposures of Pleistocene "cap rock".

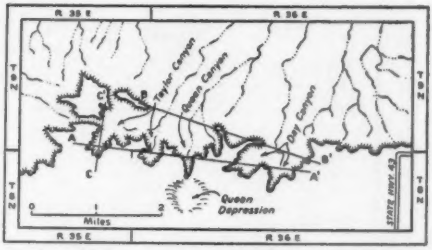

FIGURE 2.—SKETCH MAP OF TAYLOR, QUEEN AND DAY CANYONS AND QUEEN DEPRESSION
For sections *see* Figure 3.

This "cap rock" is a cemented limestone rubble approximately 2 feet thick, and largely composed of fragments of the "Pliocene cap rock". Over this rubble lies a fresh-water, platy limestone similar to that found in Puerto Canyon but not more than a foot thick.

Taylor, Queen, and Day canyons (Fig. 2), between the area covered in Plate 1 and the Texas-New Mexico line, are of particular interest because here a series of three "cap rocks", one above the other, are continuous from one canyon wall to another. These "cap rocks" also extend over a considerable area beyond the boundaries of the canyons. Multiple "cap rocks" have also been reported from Potter County, Texas (Patton, 1923), but their field relations are inadequately known.

Figure 2 of Plate 2, and the sections in Figure 3 show the relation of the "cap rocks" to each other and to the Pliocene and Triassic rocks below them. Each of the three "cap rocks" slopes toward Queen Canyon. Here also the stratigraphic interval among the three "cap rocks" is greatest, whereas at a distance they merge into a single unit, the so-called "Pliocene cap rock".

There is no apparent distinction among the three "cap rocks" except perhaps in thickness. All are composed of a platy, fresh-water limestone overlying and grading downward into a dense caliche. The upper zone of each "cap" is

a hard, dense limestone with a few isolated grains of frosted quartz. Each limestone has the same botryoidal upper surface, the same concretionary and brecciated internal structures. Each contains veins and stringers of secondary quartz. Each shows what the writer believes to be the algae described from the "Pliocene cap rock" by Elias (1931).

Obviously these are the remnants of vast depressions superimposed one upon the other and all apparently much older than any of the existing depressions. The area of the multiple "cap rocks" forms a basin with an original diameter of at least 4 miles whose center may have stood in the vicinity of what is now Queen Canyon. Much of the northern part of this basin has long since been removed by erosion, and subsequent deposition has obscured that part south of the Escarpment.

Lack of fossils prevents determination of the exact age of the several "cap rocks". Superposition, however, demonstrates the relative ages of the limestone along the Escarpment. If similar "cap rocks" crop out on the Plains then even relative ages would be impossible to establish for there exposures are poor and scattered. The inference is immediately apparent that if multiple "cap rocks" having the characteristics of the "Pliocene cap rock" are present beneath the Plains, then the "cap rock" cannot there be used as an identifiable horizon. Even if one of these "cap rocks" actually marked the end of Pliocene deposition, a supposition as yet undemonstrated, it would still be unidentifiable.

The "cap rock" along the Northern Escarpment ranges in altitude from 4920 feet north of Wilson Lake to 4700 feet east of Grapevine Canyon (Pl. 1). On the east side of Apache Canyon, the "cap rock" slopes from an altitude of 4900 to 4760 feet or a vertical interval of 140 feet in 2½ miles—a grade of 1 per cent. A similar slope exists from Deadman's Canyon southeast along the Escarpment. "Cap rock" crops out south of the Escarpment on the Plains,

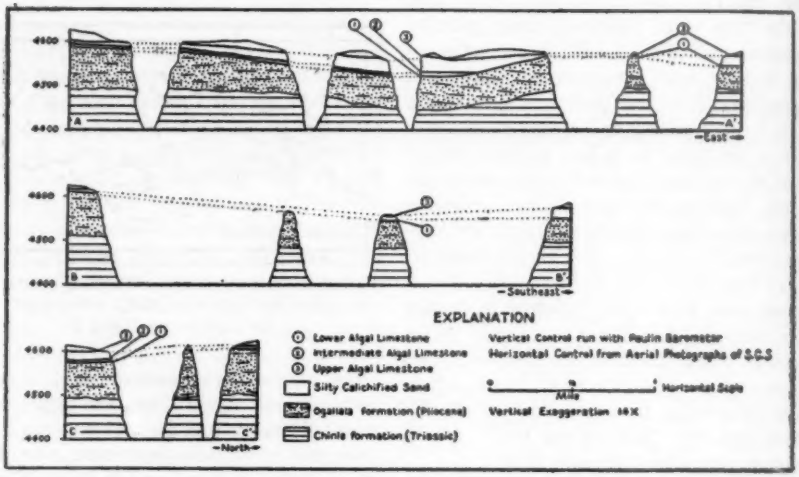

FIGURE 3.—SECTIONS TO SHOW RELATIONS OF MULTIPLE "CAP ROCKS" IN TAYLOR, QUEEN, AND DAY CANYONS

For location of sections *see* Figure 4.

FIGURE 1. CLIFF OF PLEISTOCENE "CAP ROCK" AT PUERTO CANYON
(*A*, Pl. 1) Automobile parked on "Pliocene cap rock."

FIGURE 2. MULTIPLE "CAP ROCKS" OR ALGAL LIMESTONES AT PUERTO CANYON
(1) lower "cap rock"; (2) intermediate "cap rock"; (3) higher "cap rock."

"CAP ROCKS" IN PUERTO AND QUEEN CANYONS

FIGURE 1. LEACHED (GRAY) AND UNLEACHED (WHITE) OGALLALA FORMATION AT TIPTON CANYON

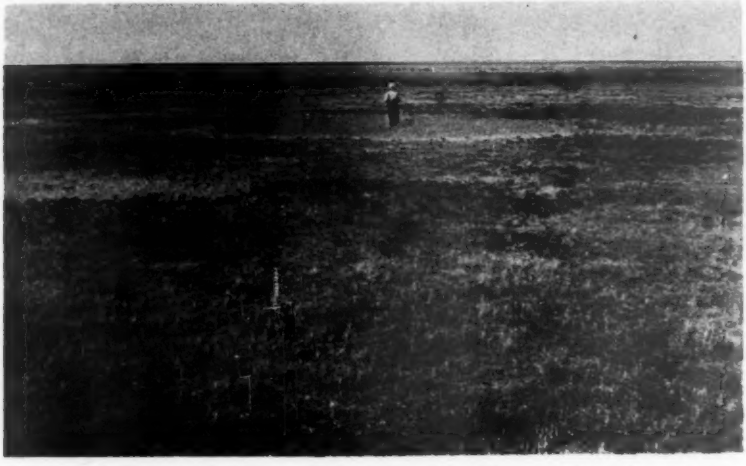

FIGURE 2. SMALL DEPRESSION, PROBABLY A TRUE "BUFFALO WALLOW"
San Jon site in background.

LEACHED AND UNLEACHED OGALLALA FORMATION AND "BUFFALO WALLOW"

chiefly around the depressions but in some instances high on the slopes of the basin. Well logs indicate that a "cap rock" is universally present beneath the "cover" of the Plains. It is missing only within the lowest portions of the depressions.

The scanty information from surface exposures and well logs indicates that the "cap rock" underlying the Plains has broad gentle undulations. The interpretation of these undulations is complicated by at least two factors: What was the original form of the "cap rock"? and, Is there a single "cap rock", or are there several?

If the formation of the "cap rock" represents the final phase of Pliocene deposition and if this "cap rock" was once a continuous sedimentary unit sloping east and southeast, then the existing undulations may be gentle folds due to post-Pliocene deformation. Such folds would have the general aspect in section as diagrammatically illustrated in Figure 4a. However, if the "cap rock" formed after Pliocene sedimentation had ceased and continued to be formed in the Pleistocene, the undulations may be original and still resemble the folds suggested in Figure 4a.

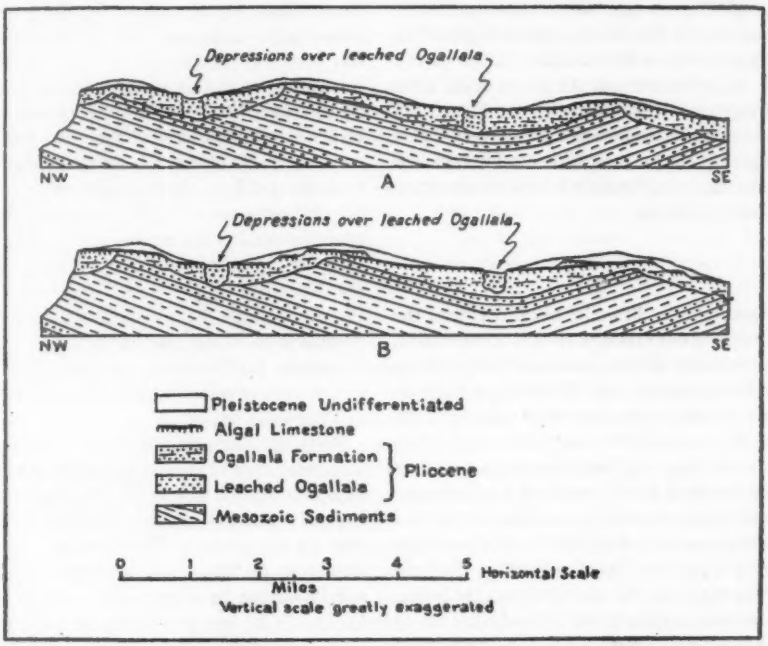

FIGURE 4.—HYPOTHETICAL SECTIONS SHOWING POSSIBLE RELATION OF THE "CAP ROCK" TO THE UNDULATIONS OF THE PLAINS

Within certain depressions are beds of limestone or "cap rocks" which are different lithologically and younger than the "Pliocene cap rock". Some of these "cap rocks", such as those exposed in Puerto and Deadman's canyons, are obviously Pleistocene. Other limestones such as the multiple "cap rocks" in Taylor, Queen,

and Day canyons are similar lithologically to the "Pliocene cap rock" and cannot be definitely assigned to the Pleistocene. At scattered and obscure exposures away from the Escarpment there is no guarantee that a series of "cap rocks" of different ages may not exist, and some of these may be Pleistocene; others could not be distinguished from the so-called "Pliocene cap rock". A section through the Plains might show the "cap rock" as illustrated in Figure 4b. Such conditions would strongly suggest that a large part of the "cap rock" is not Pliocene and that the "cap rock" cannot be used as an extensive horizon marker.

Origin of the "Cap Rock"

As set forth by Price (1933) and others, the "cap rock" is essentially a subsoil phenomenon, a concentration of lime carbonate by the leaching of overlying material. However, as much as 10 feet of almost pure limestone may form the top of the "cap rock"; thus there is too much lime in the "cap rock" unless one assumes that several hundred feet of material has been successively leached and removed. Elias (1931) has raised the question of deposition in lakes. That traces of calcareous algae are found in the "cap rock" is true, but the size of the water bodies remains in question. If the water bodies are to be lakes, then one must suppose that the "cap rock" was once a continuous smooth surface of low gradient on which lakes could exist. The existing undulations of the "cap rock" must be later. Even if the "cap rock" were formed on the slope of deposition of the Pliocene, this slope must have had a gradient of about 5 feet per mile, and the lakes must have been small and evanescent. On such a slope the ground water would have been high, and deposition of lime and other salts of evaporation would have been possible as suggested by Theis (1936). The ground water would have been replenished by local rainfall and also by flood water brought in by streams. The quantity of lime available would thus be limited only by the length of time these relations continued. The bulk of the "cap rock" might thus be deposited, and most of the calcareous cement of the Pliocene might be obtained in the same manner.

The upper part of the "cap rock" of almost pure limestone with its platy structure and botryoidal surfaces, its pisolites and its brecciated character, would be the product of secondary deposition in standing or sluggishly moving water. The nature of these water bodies is not clear. Smith (1940) and Frye (1945b) argued against the vast lake of subcontinental size originally postulated by Elias (1931) for the deposition of this algal limestone. More recently Elias (1948) has suggested smaller water bodies as an environment for "cap rock" deposition. Theis (1936) has reasoned that small pools created by a rising ground-water table in late Pliocene time would provide a suitable environment for the genesis of the platy limestone. He even suggests that the formation of the limestone might carry over into early Pleistocene time. Frye (1945b) argues that the irregularities consequent upon active alluviation across the Plains may have been flooded by a high ground-water table in the late Pliocene as alluviation ceased.

Bretz and Horberg (1949) have described the brecciated zone of "cap rock" as the uppermost zone in a caliche soil profile reflecting alternating moist and dry climates from the late Pliocene to the present. They consider that the formation of "cap rock" and the underlying caliche cannot be restricted to a single period in time or to one physiographic environment. The origin of the depressions postulated below may support this interpretation.

The question at issue is when did this upper brecciated part of the "cap rock" form? If it is Pliocene, then the undulation of the "cap rock" must be largely deformational. If this deposition occurred in the Pleistocene on an undulatory, post-Pliocene surface, the undulations would be the result of Pleistocene erosion, and no local deformation is indicated. If several "cap rocks" are involved, then a Pleistocene age for at least some "cap rock" formation is also possible and no deformation is demanded.

The undulations and the distribution of the depressions (Pl. 1) give no complete answer to this question. That must be derived from a study over a larger area and an analysis of all the pertinent facts. Nevertheless, the broad linear depressions exist. Their mere presence must play a significant role in the formation of the modern depressions.

The Depressions

Dissected Depressions

San Jon depression.—Although numerous dissected depressions were studied along the Northern Escarpment of the Llano Estacado, only the San Jon depression is discussed in detail.

A small and breached depression approximately 1 mile in diameter contains the archeological deposits of the San Jon site. The site takes its name from the small town of San Jon 10 miles to the north in the valley plains of the Canadian River. The slopes of this depression are scarcely perceptible from the near-by Plains, particularly if first seen after coming over the rugged Escarpment from the valley of the Canadian. The lip of the basin is virtually horizontal, and the only thing to catch the eye is a low and gentle sand hill rising slightly above the level on its eastern side.

Within this almost undetectable and shallow saucer, gulches 50 to 100 feet in depth have fashioned a weird clawlike pattern (Fig. 5). The ephemeral streams which cut the individual fingerlike gulches join and flow through a narrow canyon to the valley below the Escarpment. Within the depression the gulches expose a maximum of 100 feet of unconsolidated deposits. Lake and stream deposits of Pleistocene age overlie a basal, even-grained, well-sorted, buff sand. These deposits occupy a roughly circular area having a diameter of approximately 2500 feet. This is the center or lowest part of the depression, and these beds, briefly described in Table 1, form a sample of what may underlie many undrained depressions.

An eolian origin for the Basal sand is suggested by its sharp near-vertical contact with the lime-cemented sandstones of the Ogallala, by its grade size and sorting coefficients as indicated in Table 2, and by the large percentage of wind-frosted quartz grains. Close inspection, however, reveals no structures or cross-bedding such as would be expected had the material been dumped into a depression by wind. The nature of the sand remained a mystery until the discovery of similar sands in Tipton Canyon, the next canyon east. Here an unconsolidated buff sand, identical to the Basal sand of the San Jon site, exhibits all degrees of gradation into the highly calcareous sand of the Ogallala formation (Pl. 3, fig. 1). With this illuminating example in mind the inference is plain that the Basal sand at the San Jon site is also a Pliocene sand leached of its calcium carbonate cement. A mechanical analysis of the leached and unleached Ogallala at the San Jon site brings out startling similarities (Table 2). Furthermore, the unaltered Ogallala contains a high percentage of wind-frosted grains similar to those in the Basal sand.

The "cap rock", exposed only along the northern and eastern sides of the depression, appears to slope into the depression as suggested in the section of Figure 6. Part of the slope is due to thinning by erosion. One must infer that the "cap rock" is either deformed from an originally more nearly perfect plane or that it was laid down in broad undulations. Whatever the cause of the slope it is apparent that an older broad gentle depression existed before the formation of the present depression.

At the site the Ogallala is over 100 feet thick. At its base a boulder conglomerate composed chiefly of blocks of Cretaceous quartzite crops out in the canyon which now drains the depression. It overlies Upper Cretaceous shale which here dips 3° SE. The Ogallala lies within a valley cut into the underlying Cretaceous rocks. It is about 100 feet thick at the site (Fig. 6). Slightly over half a mile to the north at a point along State Highway 39 it is 40 feet thick. Eastward along the Escarpment from the site the Ogallala again thins, as the elevation of its contact with the Cretaceous increases.

Where the Ogallala sandstone has been leached, the "cap rock" is gone, and the Ogallala is leached down to, or close to, the contact with the underlying Upper Cretaceous Purgatoire formation. Some of this leached sand has been removed, and a small basin below the general level of the "cap rock" has formed within the large trough originally defined by the gently rolling "cap rock".

Within this second basin the greatest thickness of Pleistocene deposits was laid down. Breaching was accomplished comparatively recently. The Pleistocene deposits show that exterior drainage was developed before the site was occupied by the people who used the Eden Valley Yuma projectile points and hunted the

FIGURE 5.—MAP OF THE SAN JON SITE

TABLE 1.—SUMMARY OF BEDS EXPOSED AT THE SAN JON SITE

	feet
Modern deposits of the arroyos Periodically moved by flood	0 to 10
—disconformity—	
Wheatland formation	
Low terrace. Base 1 to 5 feet above grade of arroyos. Fine gravel, sand, and silt.	5 to 15
—disconformity—	
Intermediate terrace. Base 8 to 10 feet above grade of arroyos. Materials as above.	5 to 15
—disconformity—	
High terrace. Base 20 to 25 feet above grade of arroyos. Materials as above. Contains bones of *Bison bison*. No artifacts found.	10 to 15
—disconformity—	
Sand Canyon formation	
Reddish to reddish-brown alluvium. Alternating beds of sand and clayey-humic material. Contains numerous iron-manganese nodules averaging ¼ inch in diameter. Lime occurs as tubules and as films particularly along joints in the clayey horizons. Occurs in broad channels cut into underlying formations 40 to 50 feet above grade of modern arroyos. Contains artifacts ("Collateral" Yuma and Clear Fork (?)), bones of *Bison bison*, and the planorbid *Helisoma tenue sinuosum* (Bonnet), a modern mollusc common in the area.	0 to 50
—disconformity—	

TABLE 1—*Continued*

	feet
San Jon formation	0 to 50
Zone 3. Dark blue-gray clay grading laterally into greenish clay and into reddish compact sandy alluvium toward the borders of the basin. Contains nodules and plates of iron-manganese oxide and concretions and plates of calcite. Lime plates occur on laminations and on strong vertical jointing. Bones of extinct bison and artifacts are found in top of the blue clay as is the planorbid *Helisoma tenue sinuosum* (Bonnet).	0 to 45
Zone 2. Discontinuous lenses of fine gray-white volcanic ash	0 to 1.5
Zone 1. Crust of iron manganese oxides, 0 to 3 feet thick overlain by bleached white sand with clay lenses 0 to 4 feet thick; grades toward periphery of basin into reddish laminated clay with green clay seams in joints 3 to 4 feet thick, which in turn grades into reddish alluvium. Proboscidian and bison bones in the clay facies.	0 to 7
—disconformity—	
Basal sand (Decalcified Ogallala)	
Upper zone, 5 to 10 feet thick, has vertical, columnar jointing. Calcareous concretions and iron-manganese flecks and clay-filled cracks. Lower zone, 35 to 40 feet thick is a brown to buff sand without laminations. It contains vertical joints. Near the base has calcareous concretions. In most places it is separated from the unchanged Ogallala formation by vertical contacts.	40 to 50
—disconformity (hidden)—	
Lower Cretaceous Shale and Sandstone.	

modern species of bison. Furthermore, the depression was breached after the occupation of the site by those who hunted a giant and extinct bison and who fashioned the San Jon point at a time that may be equivalent with the true Folsom horizon (Roberts, 1942).

The Pleistocene deposits include lake or pond deposits and subsequent stream and terrace deposits which indicate that extensive arid periods were followed by periods somewhat moister than the present.

Other dissected depressions.—Within the area included in Plate 1 Grapevine, Deadman's, Tip-

ton, part at least of Apache, Puerto, and Moon canyons and the unnamed re-entrants labeled B and C are breached and dissected depressions. Outside the area mapped, Landrigan Canyon

TABLE 2.—PHYSICAL CONSTANTS OF BASAL SAND AND UNALTERED OGALLALA FORMATION AT THE SAN JON SITE

	Diameter in mm.			Sorting	Per cent by weight sand 1 — .05
	Q_1	M	Q_3		
Basal sand.....	0.29	0.2	0.12	1.56	97
Unaltered Ogallala.	0.25	0.17	0.12	1.483	90

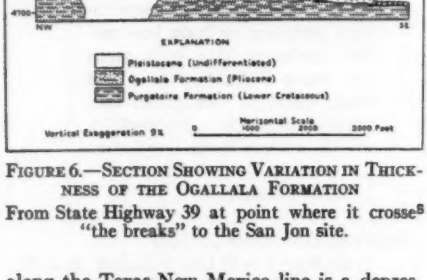

FIGURE 6.—SECTION SHOWING VARIATION IN THICKNESS OF THE OGALLALA FORMATION
From State Highway 39 at point where it crosses "the breaks" to the San Jon site.

along the Texas-New Mexico line is a depression now breached and partially excavated. Taylor, Queen, and Day canyons are similar features. Several contain Pleistocene deposits correlative at least in part with those in the San Jon depression. The age of breaching varies, but most of them have been cut into during relatively recent geologic time.

Summary statement.—Study of the interior of these depressions reveals the following: (1) There is no evidence of collapse within the Pliocene beds nor is there any evidence of collapse in the bedrock; (2) an unconsolidated sand derived directly from the Ogallala formation through leaching of its calcareous cement is present, or once was, in each depression; (3) large quantities of the leached Ogallala were removed before deposition of any Pleistocene deposits and prior to the breaching of the depressions; (4) some of these depressions lie within an older and broader depression involving the "cap rock"; and (5) several ages of "cap rock" exist. The mid- or late-Pleistocene "cap rocks" are distinguishable from the "Pliocene cap rock" although confusing at first sight. The presumably much older series of three "cap rocks" exposed in Taylor, Queen, and Day canyons are lighologically like the so-called "Pliocene cap rock". All these facts may be applied to a consideration of the undissected depressions of the Plains described below.

Undissected Depressions

Cuneva depression.—Outside the area map of Plate 1, this complex depression exhibits characteristics unique in the Southern Plains seen by the writer. The Cuneva depression, locally referred to as "The Cunevar", lies near the northern edge of the Plains about 35 miles south and west of the San Jon site (*A*, Fig. 1). The main depression is a flat-floored oval about 2 by 2¼ miles in maximum dimensions. Most of the floor is an alkaline meadow in which are three shallow alkaline lakes. This floor is surrounded on the west and south by a smooth, grassy, undissected scarp of "Pliocene cap rock". On the north and particularly on the northwest, gulches lead into the depression. An area of dissected topography as large as the major oval drains into the flat through these gulches.

Near the west end of the basin and at the foot of the north rim in the center of sec. 9, T. 7 N., R. 28 E., a sinkhole has recently formed. According to Giles Brown, a local rancher, the first collapse occurred 35 to 40 years ago, the second in 1934. The sinkhole is a doubly oval depression 150 to 300 feet wide and 40 to 60 feet deep. A dry stream enters the older and larger collapse area by a falls over Triassic rock. It runs across the floor of this depression to the conical sink of 1934. The walls of both sinks consist of Ogallala sand, feebly cemented. In addition to the outcrops at the falls, Triassic red shale crops out near the junction of the two sinks and at the base of the north wall of the later sink.

Whatever the origin of the Cuneva depression as a whole, the features of the sinkhole demonstrate collapse involving the Pliocene and continuing down into the underlying Trias-

sic. That caverns may form in the rock deep underground seems reasonable from a consideration of the log of a well drilled by the Ohio Oil Company in 1926–1928 in the SW. corner of SE¼ sec. 24, T. 7 N., R. 29 E, approximately 7 miles east of the eastern rim of the Cuneva depression, about 9½ miles from the sink, and less than 2 miles from the brink of the Escarpment. The log of this well is condensed in Table 3.

General description of the depressions.—The depressions other than the Cuneva depression show no evidence of collapse. Within the area of Plate 1 the depressions of the Plains are distributed linearly along broadly defined troughs separated by low and equally broad ridges. These depressions differ in size, shape, and depth. Many exceed a mile in diameter and are in several instances over 50 feet deep. Others, like Szaloy, Gordon, Legrone, and Wilson lakes in the northern part of the area, lie in depressions less than a contour interval in depth and less than half a mile in diameter. Still others are so shallow as to be noticeable only by a vegetative change. These latter depressions seem to be incipient. On the ridge east of the San Jon site, such depressions, 25 to 30 feet in diameter and 3 inches to 1 foot deep, occur on grassy slopes (Pl. 3, fig. 2). Small depressions of this type appear to start with a slight impedance of drainage by the grass to form a shallow pool after rain. Cows, and formerly buffalo, then trample the wet ground for a few mouthfuls of water. Between the compaction due to trampling and the carrying off of mud on their hooves or coats the depression is deepened and widened. These "Buffalo Wallows" thus suffer from "Zoogenous erosion" in Passarge's sense and are related to the Charcos (Bryan, 1925, p. 121–123) of many parts of the Southwest.

Deflation of depressions.—Many depressions have a low, sandy hill on their eastern sides. Some of these hills are prominent enough to be defined by the 20-foot contour interval (Pl. 1). A few, such as double-crested Delaney Hill east of Hatfield Lake and the hills to the east of Best, Watkins, and Owens lakes, rise slightly above the general level of the Plains. The hills east of such lakes as Wheatland are marked only by the bowing of a single contour. Several lakes have no hill to their east that is apparent on the map. However, a gentle rise is in most instances perceptible to the eye. Furthermore, the eastern side of the lake always tends to be the sandy side.

TABLE 3.—CONDENSED LOG OF OHIO OIL COMPANY WELL
SW Corner SE¼ sec. 24, T. 7 N., R. 29 E.

	Thickness	Depth
	(Feet)	(Feet)
Cover and Pliocene	205	205
"Red beds"—sandstone and shale	1535	1740
Red and gray shale interbedded with a total thickness of 133 feet of salt and gypsum	223	1963
"Red beds"	205	2168
Red and blue shale and anhydrite interbedded with a total thickness of 206 feet of salt	574	2742
Shale, limestone, anhydrite with a little salt	2462	5204

In addition to their common position with respect to a given lake all hills are similar in that they are composed largely of wind-blown sand which presumably was blown from the depressions on the west. However, little sand is now blown out of the depressions. In rainy years the lakes are too shallow to have beaches, and their margins are more or less covered with grass and weeds. In dry years some loose material is removed by wind, but the floors of most basins are baked mud. Therefore, to produce sand hills one must assume: (1) a climate so arid that the basins were almost continually dry; (2) a source of sand within the basin; (3) a weak vegetative cover incapable of holding the sand; and (4) strong dune-making winds from the west capable of moving this unanchored sand. None of these assumptions is untenable.

Recent studies of the late Pleistocene of the Southwest (Melton, 1940; Sayles and Antevs, 1941; Bryan, 1941; Huffington and Albritton, 1941; Bryan and McCann, 1943; Price, 1944) substantiate the existence in the recent past of more arid climate than now obtains. This arid period was presumably repeated after each glacial epoch. Study of the dissected depres-

sions along the Northern Escarpment of the Plains shows that the Ogallala formation is leached of its calcium carbonate cement under the depressions and thus converted from a coherent to a loose deposit. Some of this leached Ogallala sand was removed from the depressions prior to the breaching of those depressions. Wind is the only mechanism that can be invoked for this removal. Settlement within the Tertiary as a result of the removal of calcium carbonate is unlikely as the cement was deposited subsequent to the deposition of the sand itself and merely occupies the interstices between sand grains. Compaction of a well-sorted sand is negligible because most adjustment of this nature is accomplished as the sand is deposited (Athy, 1930; Fraser, 1935; Botset and Reed, 1935). Thus, using the dissected depressions as a model, one could expect deflation of those areas where the Pliocene is leached, and the consequent piling up of the sand to form hills on the eastern or lee side.

If the climate were more arid at various times in the past than it is today, water would seldom collect in the depressions. What water bodies did exist would be evanescent. Deflation would thus not be effectively curtailed nor would sediments deposited in them be thick enough or continuous enough to seal off the unconsolidated sand. Furthermore, it is not unreasonable to postulate a rainfall so ineffective that it would support only a sparse and meager vegetative cover. The winds of today are from the southwest and west and reach high velocities. There is no reason to suppose that the same was not true in the immediate geologic past, and Melton (1940) shows that the old stabilized dunes of eastern New Mexico record winds blowing in a general easterly direction.

It thus seems inescapable that the hills east of many of the depressions are the product of the deflation of those same basins. A similar origin for like features on the High Plains of Texas has been suggested by Evans and Meade (1945). One objection remains. The hills are obviously too small to account for the size of most of the depressions. One must then assume that much sand was either moved out of the area or it now forms part of the featureless "cover" above the "cap rock".

Good exposures for the study of the interior of the sand hills were available only in the eastern crest of Delaney Hill. Here a cut exposes 40 feet of sand consisting of three separate layers each terminated by a soil zone of varying maturity. Over this succession lies a thin skin of recent sand deposited during the "dust bowl" days of the 1930's and derived in part from the dry floor of Hatfield Lake and in part from plowed fields. Nothing in the oldest and thickest sand suggests an antiquity greater than late Pleistocene. If the other hills approximate this age then one must ask what happened to the sands which must have been deposited in similar forms in previous dry periods of the Pleistocene. The answer is again twofold. Either the depressions which served as sources of sand were uninitiated or were effectively sealed off from the wind, or previous sand accumulations have been leveled by time and wind. As suggested in a later paragraph and in Figure 7c the destruction of sand hills may occur during dry periods when the wind is unable to break the seal of the old lake deposits protecting the leached Ogallala. Under such conditions a sand hill would be subject to wind erosion, and no new supply of sand would be available for the growth of a new dune.

Wind deposition may affect the shape and orientation of the lakes. Deposition of eolian material on the lee side of the basins should tend to elongate the lake basins at a normal to the wind as shown by Evans and Meade (1945, Figs. 2, 3, 6, 8) in West Texas. In the area here studied there is a suggestion of elongation of certain lakes north-northwest–south-southeast, but quantitative data are insufficient to make a more definite statement.

In addition to the orientation of lakes wind deposition and erosion may cause migration of lakes into the wind. With deposition on the lee side of the basins the lake would tend to crowd into the wind. This migration is suggested by Evans and Meade (1945, Fig. 8). Again data from which definite conclusions can be drawn are lacking. Nevertheless more "cap rock" exposures are present in the southerly and westerly quadrants (Pl. 1). This suggests a crowding of the lakes into these quadrants against the "cap rock" margins of the basins. Eolian deposition would be greatest to the lee or eastern and northern quadrants during dry periods.

During high-water stages of the lakes water might clean the "cap rock" to the west and south of the thinner eolian deposits and prepare these deposits for wind transport during some later dry period.

Certainly most of the lake basins owe their origin in large part to deflation. Further, leached zones within the lime-cemented sands of the Ogallala have probably provided the foci for such wind activity. Other problems concerning the genesis of these depressions that remain, however, include the local destruction of the "cap rock", the causes of localized leaching, and the age of the formation of the depressions.

Leaching of the Ogallala

Of the many factors related to the formation of the modern depressions of the Plains the leaching of the calcium carbonate cement from the Ogallala is perhaps of most immediate importance. Marine and terrestrial beds of Pennsylvanian, Triassic, and Cretaceous age, and perhaps some beds of Jurassic age underlie the area. The distribution, character, and attitude of these beds is known from their outcrops along the Northern Escarpment and in rare valleys cut below the base of the Pliocene, and from well records. Knowledge of the pre-Pliocene bedrock of the Llano is thus limited, but the Pliocene rests on a surface of moderately low relief that has been developed across slightly deformed sandstone and shale beds chiefly of Triassic and Lower Cretaceous age. The pre-Pliocene structures as exposed north of the Llano in the valley plains of the Canadian River show a regional northeasterly trend (Dobrovolny and Summerson, 1946). The strike of the pre-Pliocene sediments is similar at various points along the Escarpment (Pl. 1). It is not unreasonable to suppose that this structural trend is likewise present beneath the Pliocene of the Plains.

An examination of the Northern Escarpment indicates that pre-Pliocene valleys have this general trend. The immediately apparent suggestion is that these are strike valleys controlled by the outcroppings of gently folded and bevelled rocks of differing resistance. Such a condition is postulated in the hypothetical cross-sections (Fig. 4). Unfortunately, there is little concrete evidence to support this supposition. The records of well logs reproduced in Table 4 contain no usable information concerning this pre-Pliocene surface. Despite the inconclusive nature of the evidence possibly the Pliocene has obscured an old trellis drainage and today the thickest deposits of the Ogallala occupy the axes of these buried valleys.

An examination of the Tertiary along the Escarpment indicates that leaching of the Ogallala is confined to the areas of its greatest thickness. Conversely, where the Tertiary is thin and the pre-Tertiary bedrock is proportionately high, there is no leaching. Is it then possible that the mere presence of a surface depression indicates a considerable thickness of the Ogallala? And does the lineation of the depressions indicate a zone of maximum thickness and the trace of a pre-Pliocene drainage? These are tantalizing questions, but they cannot be answered on the basis of present information.

We can gain some information concerning the mechanism of leaching from an examination of the well logs (Table 4). Information on many wells is lacking because the original owners and the drillers have moved away.

The thickness of the post-Pliocene "cover", where known, ranges from 0 to 40 feet. The thickness of the Pliocene is indeterminate in most logs. In well 12 it is probably close to 100 feet thick. In wells 30, 31, and 33 the Pliocene may have a similar thickness. Where known along the Escarpment the Pliocene rarely reaches 200 feet in thickness. Therefore, probably most, if not all, of the wells strike water below the base of the Tertiary.

The single most interesting fact brought out by the well logs is the relation of strong wells and weak or dry wells to the broad troughs in which the depressions lie. Water can be obtained from wells along the trough between Apache Canyon and the San Jon site. Although the wells are not overly strong this trough marks a much more favorable place to drill than does the broad ridge to the south separating this trough from the one next south. On this ridge wells 4, 5, and 6 are weak, and holes were drilled in the area of well 4 before water was struck. Wells 7, 8, 9, and 10 all mark dry holes or a series of dry holes. Water is again plentiful in the trough that runs through Wheatland.

TABLE 4.—DATA ON WELLS ON THE HIGH PLAINS IN VICINITY OF THE SAN JON SITE
(Numbers refer to Plate 1. All measurements in feet.)

No.	Owner	Surface Elevation	Depth	Thickness of post-pliocene	Depth to water	Elevation water table	Remarks
1	Gordon	4915	285	40	200+	4715+	
2	Legrone	4890	150	40	140	4750	Only strong well of 7 drilled.
3	Mullinex	4875	170?	0	160	4715	
4	Szaloy	4918	350	8	250	4668	"Red beds" at 350; Pliocene and Cretaceous equals 342.
5	Legrone	4880	?	?	218	4662	
6	Forsthofer	4900	?	?	250	4650	"Red beds" not reported, probably in Cretaceous.
7	Forsthofer	4860	590	30	dry		Ends in "Red beds".
8	?	4840–4900	300–350	20–40	dry		Six dry holes all ending in "Red beds".
9	McCagney	4805	500	25	dry		"Red beds" at 200.
10	Tillman	4775	over 200	15–20	dry		Ten dry holes in secs. 3, 10, 15, T. 8 N., R. 34 E., "Red beds" about 200.
11	Chapman	4770	?	?	180	4590	
12	Kauffman	4770	?	?	280	4490	Clay at 100 probably equals base of Pliocene.
13	Wheatland School Well	4790	270	30	260	4530	Red clay at 270, water along top of clay.
14	McCagney	4770	290	30?	270	4500	Top of Pliocene about 4740 around Wheatland Lake.
15	L. Frost	4770	250	20	245	4525	Wells 15–19 get water off top of "Red Beds" and probably within Cretaceous.
16	F. Frost	4760	225	20	220	4540	As above.
17	Swift	4760	225	20	220	4540	As above.
18	H. Frost	4730	220	18	190–220	4540	As above
19	F. Frost	4720	?	?	220	4500	As above.
20	Sanderson	4730	437	40	130	4600	Water in gravel over "Red beds".
21	Sanderson	4700	?	?	dry		Three dry holes.
22	Bryant	4710	225	20	215	4495	Probably in Cretaceous.
23	Bryant	4720	225	20	215	4505	Probably in Cretaceous.
24	?	4690	230	18	212	4378	Probably ends in Cretaceous clay.
25	Grout	4680	230	?	215?	4465?	
26	?	4670	200±	?	200±	4470±	Weak. Several dry holes in area.
27	?	4670	?	?	dry		Area of several dry holes ending in "Red beds".
28	Hatfield	4640	180	0?	170	4470	Water in sand—probably Cretaceous strong well.
29	Hatfield	4630	160?	?	150	4480	Strong well.
30	Waterfield	4710	107	?	88	4622	Water in "quick sand". "Red beds" at −107.
31	Waterfield	4710	100–110	?	90–100	4610–4620	Three wells in "quick sand".
32	?	4730	230	?	dry		Ends in "Red beds".
33	Carter	4720	110	0?	100	4620	Water in "quick sand".

Along the next broad ridge south it is difficult to obtain a reliable source of water, and several dry holes were drilled in the area of wells 26 and 27. The area marked 21 indicates three dry holes. However, well 20 is considered a good well in an otherwise unpromising area. Within the trough that runs through and includes Hatfield Lake, water is found without much difficulty.

Thus adequate supplies of ground water are concentrated along the troughs, but not along intervening ridges. This distribution of ground water must have some bearing on the leaching of the Ogallala. Where there is the greatest flow of ground water there should also be the greatest solution of cement within the Pliocene. That most wells draw their water from some horizon below the Pliocene indicates that water can escape into the lower strata, and the dissolved salts can be removed from the Ogallala formation.

The "cap rock" offers a considerable but not insuperable barrier to the downward passage of rainwater. The upper zone of the "cap rock" is a dense limestone composed of innumerable plates so that the contacts between these plates are planes of weakness. Also each plate tends to break into nodules. Incipient breaks or fractures through the "cap rock" would be quickly widened allowing the downward passage of water and dissolved salts. The downward drainage of water through the Ogallala would leach the lime cement from this formation directly below the break in the "cap rock". The zone of leaching would theoretically be controlled by the size and shape of an ever-widening hole in the "cap rock".

This method of localized leaching of the Ogallala is suggested by the conditions at the San Jon site. The contact between the leached and unleached Ogallala is nearly vertical. Around the northern and eastern sides this contact coincides with the edge of the "cap rock". It may be inferred here that the extent of the "cap rock" determines the areal extent of the unleached Ogallala. Where the "cap rock" is absent the Ogallala has been leached of its lime and converted to an unconsolidated deposit. The same relations between "cap rock" and unleached Ogallala and the absence of "cap rock" and the presence of leached Ogallala is present in the other dissected depressions examined along the Escarpment. The original breaks in the "cap rock" would tend to be localized in the troughs between ridges, as rain water would concentrate here. Thus the localization of the depressions again leads to the unsolved origin of the troughs.

Climate and Evolution of the Depressions

Processes stemming from changes of climate have been invoked to assist in explaining the formation of the depressions. The importance of these fluctuations in the process of depression development warrants their further evaluation.

During the Pleistocene southwestern United States underwent more or less rhythmic fluctuations of climate which varied from extreme aridity to a climate moister and probably somewhat cooler than that of the present. Our information is confined largely to the late Pleistocene, and the general picture of the whole Pleistocene must be sketched by extrapolation from this fragmentary knowledge. Evidence, however, marshalled by numerous workers points to a correlation between the climate of the Southwest and the glaciations and deglaciations of Canada, northern United States, and the western mountains. With the advance of ice in the North and in the higher mountains the arid and semi-arid zones of the Southwest were displaced southward and supplanted by moister and somewhat cooler climates, the "pluvials" of Antevs (1935; 1938). Winters were more rigorous, and summers cooler. Conversely, temperatures increased, and effective rainfall decreased with periods of deglaciation.

This succession of dry and moist climates with accompanying changes in process are directly applicable to the formation of the depressions. The breaching of the "cap rock" and the subsequent solution of the underlying Ogallala sands demand moisture. Thus the solution of lime carbonate and its transportation into the underground would proceed most rapidly during the moist periods correlated with the glacial advances to the north. Conversely deflation would be most effective in the arid periods correlated with the periods of ice retreat and deglaciation.

Figure 7 suggests the relation of form and process to the periods of aridity and moisture and indicates some of the variations in form which might be expected. Profile A pictures two areas in which the "caprock" has already been destroyed and the Ogallala sandstone partially leached during a preceding moist period. During a dry period wind moved the leached and unconsolidated sands to the eastern sides of the incipient basins where the coarser fraction collected as a sand dune. Many of the fines were presumably carried beyond the point of dune deposition and either removed from the area or added to the featureless "cover" of the Plains. During the subsequent wet periods the depressions filled with water and deflation ceased as indicated in profile B. Lake sediments collected in semi-permanent lakes in the basins. Further leaching of the Ogallala was hastened by greater amounts of water. Vegetation stabilized the dunes, and a soil developed. At the same time rainwash modified the original form of the dune. With the onset of a second dry period (Profile C) the semipermanent lakes disappeared, the vegetation was weakened, and deflation again predominated. In basin 1 the wind removed the previously deposited lake deposits and exposed the leached Ogallala. Eolian material was added to the remnants of the old dune east of the

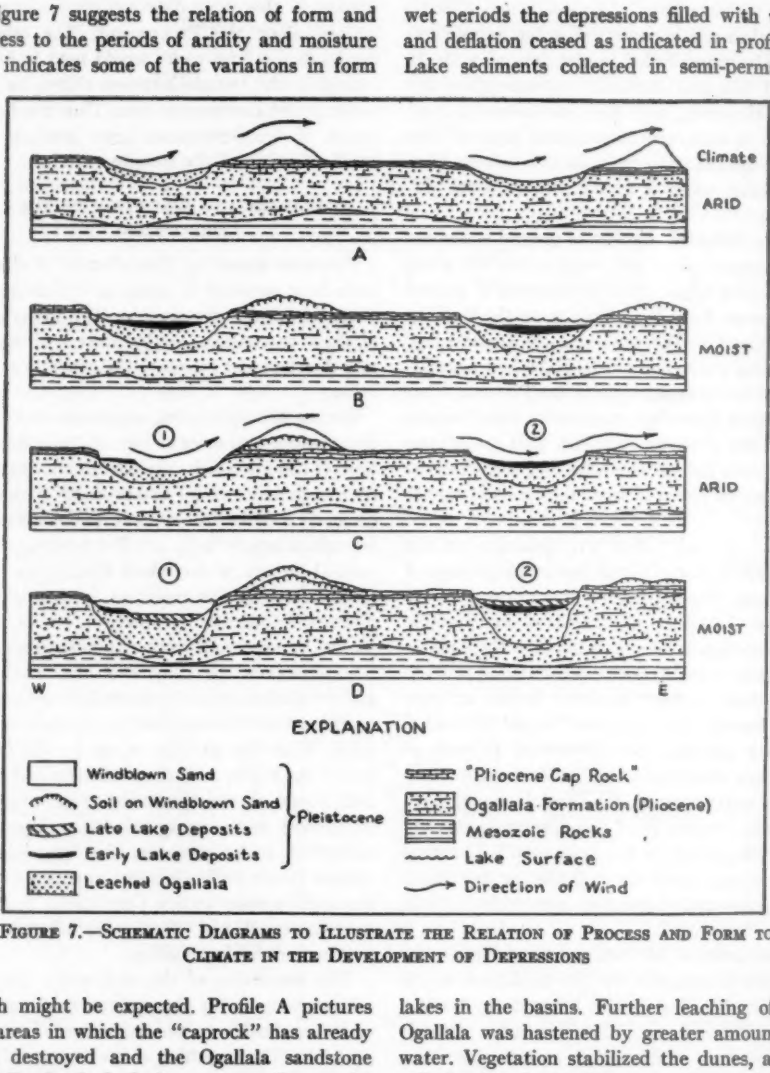

FIGURE 7.—SCHEMATIC DIAGRAMS TO ILLUSTRATE THE RELATION OF PROCESS AND FORM TO CLIMATE IN THE DEVELOPMENT OF DEPRESSIONS

depression. Small remnants of the old lake beds may have been preserved. By this time, increased removal of sand caused slumping around the edges of the basin, and the "cap rock" may thus have developed an exaggerated dip into the basin. Basin 2 of profile C illustrates other possibilities. In this instance the wind is considered to have been incapable of breaching the exposed lake deposits. No new source of sand was then available to increase the dune east of the basin. Under such conditions the basin ceased to deepen despite the arid conditions. Consequently the old dune may have been partially or almost completely destroyed by the wind.

A return to moist conditions repeated the events of the earlier moist period. In basin 1 new lake deposits formed within the recently excavated part of the basin. The total amount of new sediments may not have filled this hole and the older deposits stood as a low terrace. The dune would have been modified as previously. In basin 2 the new deposits obscured the older deposits. In both basins leaching of the Ogallala again quickened.

Furthermore, it is reasonable to suppose that erosion by the activities of animals has removed some material from the basins, particularly during the moister periods when deflation was not active and when both mud and large animals were common. Possibly "zoogenous erosion" played an important role in the initiation of a small depression, and, once started, this depression may have grown as outlined above. Nevertheless, the total amount of material removed by animals from a depression must have been relatively small, and the process can be considered of only very minor importance in the formation of the depressions.

Age of the Depressions

The depressions are the products of the Pleistocene. There is no *a priori* reason to suppose that the formation of the depressions would be favored at any one time during the Pleistocene. Two lines of evidence tend to support this assumption: (1) the age of the deposits within the depressions, and (2) the size of the depressions.

In the area studied most of the deposits are of proved late Pleistocene age or appear to be late Pleistocene or slightly earlier. Only in Taylor, Queen, and Day canyons is a greater age indicated. However, exposures in the unbreached depressions on the Plains are poor and deposits older than the late Pleistocene are to be expected within at least some of them.

Deposits of early and middle Pleistocene age are known outside the area. In western Texas along the eastern escarpment of the Llano Estacado, Evans and Meade (1945) have studied the Blanco, Tule, and Tahoka deposits of early, middle, and late Pleistocene age, respectively. They demonstrate that these beds are all lacustrine and laid down in basins developed within the Ogallala formation. The older basins are almost choked with deposits, and their extent, nature, and, in some instances, presence are known only because of recent dissection by streams. These authors conclude that "basins similar to those occupied by modern playas have been forming and filling on the Texas plains throughout most if not all of Quaternary time."

The varying diameters and depths of the modern depressions indicate that the initiation and expansion of the depressions cannot be confined to any specific horizon in the Pleistocene. In general it takes more time to develop large depressions than small ones, despite the fact that any two depressions initiated at the same time may increase in size at differing rates as suggested in the discussion of Figure 7.

It is entirely reasonable that depressions have been forming on the plains since the end of the Pliocene. Some have ceased to grow and have become so choked with deposits that there may be little recognizable surface expression. Were the surficial cover stripped from the Plains, the resulting surface should be literally pock-marked with the open scars of modern depressions and the healed or partially healed scars of "extinct" depressions. The depression is not only characteristic of the Plains of the present, but also of the past back to the beginning of the Pleistocene.

Conclusions

The shallow depressions of the High Plains of eastern New Mexico are formed not by col-

lapse into the underground or by differential subsidence in the Tertiary deposits, but by a combination of leaching and wind deflation throughout Pleistocene time.

The depressions are concentrated along the bottoms of broad shallow troughs, undulations which involve the "cap rock". A basin was initiated by the local destruction of the "cap rock" and leaching of the lime carbonate cement on the underlying Ogallala formation during a moist period of the Pleistocene. During dry periods of the Pleistocene strong winds, decreased effectiveness of precipitation, and a weakened vegetative cover resulted in the deflation of these locally leached areas. During moist periods lake deposits of clay and silt have accumulated in the depressions.

Some field evidence indicates that the so-called "Pliocene cap rock" involved in the broad troughs containing the depressions is not a continuous time-rock unit. Nevertheless data are insufficient to determine whether the troughs are the result of (1) deformation of an originally almost horizontal plate of "cap rock"; (2) depositional structures of a continuous stratum of "cap rock"; or (3) a series of different "cap rocks" of Pliocene and Pleistocene age formed at varying elevations and later largely obscured by wind-deposited material.

Field facts indicate that the depressions are developed where the Ogallala formation is the thickest, and along lines paralleling the inferred strike of the underlying, gently folded and eroded Mesozoic rocks. Do, then, the depressions mark the trace of a buried pre-Pliocene trellis drainage? Once again data are insufficient for a definitive answer.

That collapse depressions can occur in the area is shown by the Cuneva depression in which collapse involving Pliocene and Triassic beds has occurred twice within the last 40 years.

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